Volume 12, Number 5
September–October 2004

ISSN: 0869-5938

STRATIGRAPHY AND GEOLOGICAL CORRELATION

English Translation of *Stratigrafiya. Geologicheskaya Korrelyatsiya*

Editor-in-Chief
Boris S. Sokolov

http://www.maik.ru

Journal of the Fundamental and Applied Aspects of Stratigraphy and the Correlation of Geologic Events and Processes in Time and Space

МАИК "НАУКА/INTERPERIODICA" PUBLISHING

Stratigraphy and Geological Correlation

(Stratigrafiya. Geologicheskaya Korrelyatsiya)

ISSN: 0869-5938

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Stratigraphy and Geological Correlation is abstracted and/or indexed in *Petroleum Abstracts*, *Research Alert*, *SciSearch*, *Current Contents* (*Physical, Chemical*, and *Earth Sciences*) and *World Translations Index*.

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PUBLISHER

INTERNATIONAL ACADEMIC PUBLISHING
COMPANY "NAUKA/INTERPERIODICA"
PROFSOYUZNAYA UL. 90, MOSCOW, 117997 RUSSIA
PHONE: 7 (095) 336-1600; FAX: 7 (095) 336-0666
E-mail: compmg@maik.ru
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SUBSCRIPTION PRICES (2004, Volume 12, 6 issues)

US $1365.00 (North America), $1564.00 (elsewhere). Subscriptions may be entered at any time. Orders may be placed at agencies or directly. Cancellations not allowed except for duplicate payment. Prepayment required. Advance payment required in US$/US bank. MasterCard/Visa accepted. Make checks payable to *Stratigraphy and Geological Correlation* or Interperiodica.

Stratigraphy and Geological Correlation (ISSN: 0869-5938) is published bimonthly for US $1365.00 (North America), outside North America for $1564.00 per year by MAIK "Nauka/Interperiodica", 5724 Hwy. 280 East, Birmingham, Alabama 35242-5724 USA. Periodical postage paid at Birmingham, Alabama, and additional mailing offices. POSTMASTER: Send address changes to *Stratigraphy and Geological Correlation*, P.O. Box 1831, Birmingham, Alabama 35201 USA.

The journal was founded in 1993.

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English Translation Copyright © 2004 by MAIK "Nauka/Interperiodica" (Russia).

Printed in the USA by Pleiades Publishing, Inc.

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The Yudoma Group of Stratotype Area: C-isotope Chemostratigraphic Correlations and Yudomian–Vendian Relation

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Received November 25, 2003

Abstract—Representative C-isotope characteristics are first obtained for the Lower Yudomian carbonates (Yukanda Formation) from the Belaya and Sakhara sections in the Yudomian stratotype area. The least altered carbonate samples that elucidate further C-isotope chemostratigraphy of formerly studied Upper Yudomian interval were selected based on strict geochemical criteria (Mn/Sr ≤ 4, Fe/Sr ≤ 10.0, $\delta^{18}O > -10‰$ in limestones and Mn/Sr ≤ 6, Fe/Sr ≤ 15, $\delta^{18}O > -10‰$ in dolostones). In the studied samples, $\delta^{13}C$ values demonstrate the in-phase variations of comparable amplitude at the correlative levels of both sections. They are used to plot the generalized curve of $\delta^{13}C$ variations in the Lower Yudomian succession that is divisible, based on respective trends, into intervals corresponding to three subformations of the Yukanda Formation. Characteristic of the Yalan Subformation, the lower one, are prevailing high positive $\delta^{13}C$ values. In the Mal Subformation (middle one), this parameter declines down to moderate positive values with two short negative excursions complicating the general trend. Low negative $\delta^{13}C$ values are typical of the Tokur Subformation that is upper in the succession. By means of C-isotope chemostratigraphy, biostratigraphy and isotopic-geochronological data, the Yudomian type sections are correlated with terminal Precambrian successions of North Siberia, Namibia, and Oman, and the resulting inferences are as follows. (1) The pre-Nemakit-Daldynian intervals of Vendian successions in Siberia are compatible in their C-isotope characteristics with the reference chemostratigraphic curve (Saylor *et al.*, 1998) and add to it from below the interval with high positive $\delta^{13}C$ values, the lower boundary of which is at the level of inversion from positive to negative values in the cap carbonates of the lower Varangerian tillites. (2) In the stratotype area, the Lower and Upper Yudomian deposits are separated by the regional hiatus (time span of 4 to 4.5 m.y.). (3) Judging from $\delta^{13}C$ values, basal horizons of Yudomian carbonates are stratigraphically higher than the cap dolomites of the lower Varangerian tillites, and consequently the Yudomian carbonate succession is less than Vendian in stratigraphic range. (4) The direct evidence of age relation between the Vendian and basal Yudomian terrigenous deposits is unknown, but indirect considerations suggest that the latter accumulated during the Early Vendian deglaciation. (5) Negative excursions of $\delta^{13}C$ values, which are recorded in the Mal Formation of Uchur-Maya region and near the Kufai Formation top in Oman, presumably reflect events related to the late Varangerian glaciation.

Key words: Yudomian, Vendian, Varangerian glaciation, C-isotope chemostratigraphy, Uchur–Maya region, North Siberia, Namibia, Oman.

INTRODUCTION

It is a common practice to consider the Yudoma Group (Yudomian) of stratotype area (the Uchur–Maya region of southeastern Siberia) as an equivalent of Vendian, and respective units have been described at times under the last name, although there is no data, as stressed by Sokolov (1967), that lower boundaries of both subdivisions are concurrent. That remark made almost 40 years ago is actual even now, and three viewpoints on the Yudomian–Vendian age relation were under debates during the time past. In some publications of the 1960s to 1970s, it was argued for that the Yudomian corresponds not only to the Vendian of East European platform, but also to the uppermost Riphean horizons of the Uralian type sections (Keller and Semikhatov, 1968; Semikhatov *et al.*, 1970; Keller *et al.*, 1974). Somewhat later, it was suggested that the Yudomian corresponds exclusively to the upper, post-tillite part of the Vendian (Semikhatov and Serebryakov, 1983; Semikhatov, 1991). In general, the idea of Vendian–Yudomian synchronism was dominant since the recognition of the latter (Zhuravleva and Komar, 1962) until present days (Sokolov, 1967, 1979, 1997; Kho-

mentovskii, 1975, 1976, 1984; Semikhatov and Serebryakov, 1983; Sokolov and Fedonkin, 1985, 1990; Khomentovskii and Karlova, 1994a, 1994b, 2002, and references therein), and this situation is acknowledged in the Precambrian stratigraphic scale of the USSR (Semikhatov *et al.*, 1991).

The first viewpoint was based on assumption that microphytoliths characteristic of the Yudoma Group appear concurrently in different regions, and the Yudomian lower boundary was displaced at the pre-Vendian level in view of their occurrence below the Laplandian (Varangerian) tillites in European areas of Russia. Soon it became clear that microphytolith assemblages are diachronous (Semikhatov, 1974; Khomentovskii, 1976, 1984; Yakshin, 1979, 1985), and the idea lost substantiation. Adherents of the second viewpoint correlated the pre-Yudoma unconformity with the glacioeustatic sea-level drop at the time of Varangerian glaciation. Finally, those who argued for the third viewpoint referred to Ediacara fauna present in middle horizons of Vendian and Yudomian, to available radioisotopic dates, and to a similar structural position of deposits under consideration in the East European and Siberian platforms, what was a basic argument. Nevertheless, geological observations and data of C-isotope chemostratigraphy showed that basal horizons of the Yudomian are not identical in age in different areas of Siberian platform (Semikhatov *et al.*, 1970; Khomentovskii, 1976; Starnikov *et al.*, 1983; Davydov and Sobolev, 1990; Kaufman *et al.*, 1996; Pelechaty *et al.*, 1996a). These results lowered significance of the last argument, while localization of Ediacara fauna meant nothing about time-relations of the Vendian and Yudomian lower boundaries. As for the mentioned radioisotopic dates, they appear to be invalid in terms of present-day analytical requirements, although there is another opinion (Khomentovskii, 2000).

There are two moments, which delayed for a long period the resolution of problem of Vendian–Yudomian relation with the help of C-isotope chemostratigraphy. First, C-isotope parameters of the Yudoma Group stratotype in the Uchur–Maya region were known for carbonates of its upper formation only, which was studied in western sections nearby the Aldan River (Magaritz *et al.*, 1986; Brasier *et al.*, 1993, 1994b, and other works), whereas only a few determinations of $\delta^{13}C$ values were published for rocks of the lower formation sampled in distant sections (Pokrovskii, 1996; Podkovyrov *et al.*, 1996). Second, the Vendian type sections are composed of terrigenous sediments, and C-isotope characteristics of organic carbon have been studied in their upper (<551 ± 4 Ma) interval only (Strauss *et al.*, 1997). A possibility to overcome these obstacles appeared just recently, when representative C-isotope and other data necessary to control chemostratigraphic correlations were obtained for the terminal Precambrian in several regions. To realize this possibility, we put forward three objectives in this work: (1) to determine $\delta^{13}C$ variations in most thick and complete sections of the Lower Yudomian deposits; (2) to compare C-isotope parameters in the Yudomian type succession and sections of the terminal Precambrian in other regions; (3) to substantiate telecorrelation between the sections in question and to clarify the relative position of the Yudomian and Vendian basal levels. In this work, the Cambrian lower boundary is accepted to be at the Tommotian Stage base, i.e., at the base of its first *N. sunnaginicus* Zone, which is traditional for Russian stratigraphers (*Stratigraphic Code*, 1992; Rozanov *et al.*, 1997). According to isotopic dating, this boundary is 535 ± 1 Ma old (Rublev, 2000; Semikhatov, 2000). Abroad the Cambrian lower boundary is placed much lower. According to resolution of the International Stratigraphic Commission, which is approved by International Union of Geological Sciences in 1992, it is defined in the type section of Burin Peninsula, southeastern Newfoundland, at the appearance level of trace fossils *Trichophycus (= Phycodes) pedum*, i.e., at the level separating the Cambrian *T. pedum* Zone from underlying *Harlaniella podolica* Zone of ichnofossils. The isotopic age accepted for this level is 543 Ma (Brasier *et al.*, 1994a; Landing, 1994).

GEOLOGICAL BACKGROUND, AGE CONSTRAINTS AND STRUCTURE OF THE STUDIED SECTIONS

In the Uchur–Maya region, the Nel'kan thrust fault of submeridional strike (Fig. 1) separates the Uchur–Maya plate with a subhorizontal attitude of Riphean, Vendian, and Cambrian strata, from the easterly Yudoma–Maya belt, where the same beds are thicker, partly change their lithologic composition and form large open folds that are locally faulted (Nuzhnov, 1967; Semikhatov and Serebryakov, 1983; Khudoley *et al.*, 2001). Rocks of the Yudoma Group divided in two regional formations are widespread in both structural zones. The Aim Formation of the plate, which is the lower one, was termed at times within the belt as the Yukanda Formation. In the plate sections, it unconformably overlies different older deposits, from Archean to Upper Riphean, and intrusions of central type, which crosscut all the Riphean strata (Semikhatov and Serebryakov, 1983; Figs. 41 and 44). In the Yudoma–Maya belt, the Lower Yudomian deposits are missed from sections of western and southern marginal zones, being represented by considerable sedimentary successions in the east and northeast, where respective sediments overlie the Upper Riphean deposits and associated basaltic flows and sills (Sukhorukov, 1984). The upper Ust-Yudoma Formation of the group is alternatively termed the Sardana Formation in the belt, where it rests on the eroded surface of the Lower Yudomian and Upper to Middle Riphean deposits (Davydov and Sobolev, 1990; Fig. 3). Almost everywhere in the plate, the Ust-Yudoma Formation overlies the erosion surface of the Aim Formation and overlaps the plate basement in the west. The upper limit of the Yudoma Group cor-

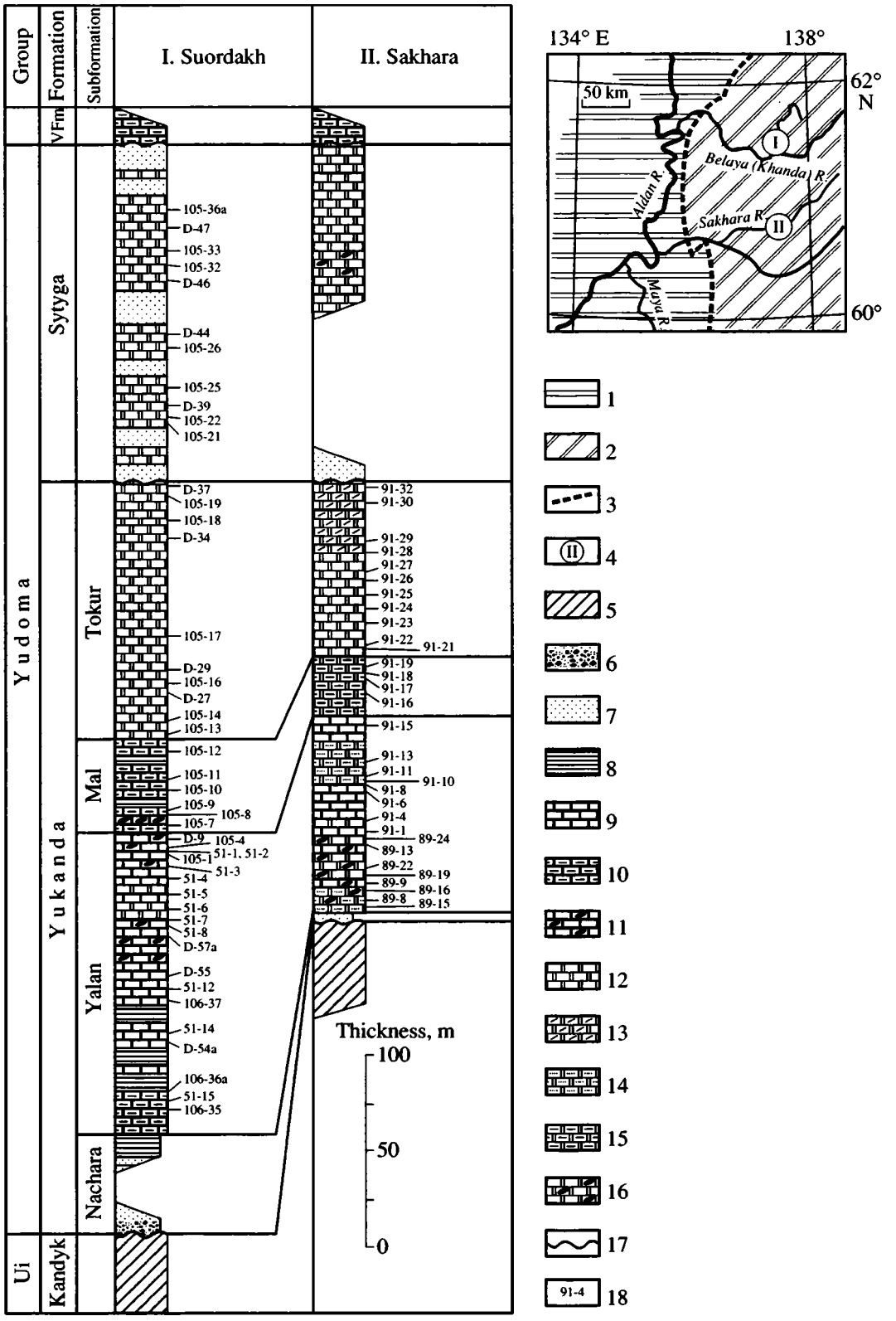

Fig. 1. Schematic geology of the region (inset map) and lithostratigraphy of the studied Yudoma Group sections: (1) Uchur–Maya plate; (2) Yudoma–Maya belt; (3) Nel'kan thrust fault; (4) Suordakh (I) and Sakhara (II) sections of the Yudoma Group; (5) shales and siltstones, Kandyk Formation; (6) conglomerates and gravelstones; (7) sandstones; (8) shales; (9) limestones, varieties enriched in organic matter included; (10) clayey limestones; (11) limestones with cherty lenses; (12) massive dolostones; (13) cross-laminates dolostones with grain stone interlayers; (14) sandy dolostones; (15) clayey dolostones; (16) dolostones with cherty lenses; (17) unconformities; (18) sample no. and level of sampling; (VFm) Variegate Formation of the Lower Cambrian.

responds to the base (*N. sunnaginicus* Zone) of the Variegate Formation of the Lower Cambrian Tommotian Stage. In the belt, there is recorded a hiatus and, locally, angular unconformity underneath the last formation (Starnikov *et al.*, 1983; Sukhorukov, 1984; Davydov and Sobolev, 1990, Khomentovskii and Karlova, 1994a), whereas in the plate it overlies the erosion surface of Ust-Yudoma deposits (Semikhatov and Serebryakov, 1983; Magaritz *et al.*, 1986). The time span of hiatus below the Variegate Formation is variably estimated (e.g., Khomentovskii, 1998, vs. Knoll *et al.*, 1995b), although chemostratigraphic data suggest that it was quite long (Knoll *et al.*, 1995b; Kaufman *et al.*, 1996; Kouchinsky *et al.*, 2001).

Paleontological characterization of Yudoma Group. Paleontological remains are most frequent in the Ust-Yudoma Formation. In the Uchur–Maya plate sections, three upper member of this formation yielded successive assemblages of small shelly fossils (SSF) characteristic of the *A. trisulcatus* and *P. antiqua* zones of the terminal Nemakit-Daldynian Stage of the Vendian. In the Yudoma–Maya belt, representatives of above assemblages were found in the middle and upper intervals of this formation (Semikhatov *et al.*, 1970; Shenfil', 1991), and at the level of 110 m below its top, taxa of the lower assemblage occur in association with mineralized *Namacalathus* remains and *Cloudina* shells (Zhuravlev *et al.*, in press), which are known in the Vendian above tillite beds in many regions (Grant, 1990; Burns *et al.*, 1994; Saylor *et al.*, 1998; Brasier *et al.*, 2000; Conway Morris *et al.*, 1990). In addition, soft-bodied Metazoa *Cyclomedusa* ex gr. *plana* and *Medusinites* were found in association with endemic *Suvorovella* and *Majaella* of unclear biological affinity near the eastern border of the Uchur–Maya plate (Khomentovskii and Karlova, 1994a, 1994b, 2002). Representatives of two former genera are known from different Upper Vendian horizons of several regions, and Fedonkin reported on typical *Cyclomedusa plana* found in the eastern White Sea region above the tuff interlayer that is 555.3 ± 0.3 Ma old according to U–Pb age determination (Martin *et al.*, 2000).

Morphologically simple filamentous and coccoidal silicified microfossils rather abundant in the Yudoma Group are useless in stratigraphic aspect (Lo, 1980; Pyatiletov, 1988; Sergeev, 2002). Rare acanthomorphic acritarchs *Micrhystridium* sp., problematic *Megahystrichospaeridium*, and remains of true protists *Osculoshpaera tarynnachia* and cyanobacteria *Obruchevella* sp. are of a broad stratigraphic range, being characteristic of the Upper Riphean and Vendian. Organic-walled microfossils from Yudoma deposits are represented by either transit morphotypes, or Siberian taxa occurring in the Vendian–Cambrian boundary horizons and partly at deeper levels. The Yudomian stromatolites are usually confined to analogues of the Yudoma Group, and some of them occur in the terminal Upper Riphean outside Siberia (Semikhatov *et al.*, 1970; Semikhatov and Serebryakov, 1983).

Age constraints. For a long period, it was believed, based on isotopic dates for Yudomian glauconites and pre-Yudomian intrusions of central type, that the group ranges in age between 650 ± 20 and 570 (530) Ma or between 620 ± 20 and 580 ± 20 Ma (see Keller and Semikhatov, 1990; Khomentovskii, 2000 for review and references). These assessments cannot be regarded as reliable now. The U–Pb dates obtained for intrusions are highly discordant, ranging from 650 ± 100 to 600 ± 40 Ma, and the common Pb content in minerals used for age determination is abnormally high. K–Ar systems in biotite from outer near-contact zones of intrusions and in phlogopite from comagmatic kimberlites were definitely disturbed as follows from the great range of age values obtained (740–586 Ma). K–Ar dates for mineralogically unstudied glauconites from the Yudoma Group and Variegate Formation base, which were published over 30 years ago range from 660 to 560 Ma and from 610 to 510 Ma, respectively. At the same time, monomineral fractions of Al-glauconite from that formation base, which are properly characterized in terms of composition and structure, yielded K–Ar dated of 476–454 Ma (Semikhatov *et al.*, 1987). It is logical to think in view of these dates that mineralogically unstudied glauconites had disturbed K–Ar systems and were contaminated by a clastic K-bearing substance. Similarly unreliable is the Rb–Sr errorchron date of 639 ± 30 Ma obtained for a fine (<1 μm) clay fraction from the Upper Yudomian (incurrent understanding) shales (Krasnobaev *et al.*, 1986). As is known (Gorokhov, 1996), fractions of this kind consist of noncogenetic clay minerals retaining isotopic memory about different stages of lithogenesis.

The Pb–Pb isochron date of 553 ± 23 Ma, which has been obtained for limestones sampled from basal interval of the Ust-Yudoma Formation in middle courses of the Yudoma River (Semikhatov *et al.*, 2003), appears to be correct in methodic aspect. This date defines the time of early diagenesis and is compatible, with due account for stratigraphic position of studied samples, with U–Pb date of 543.9 ± 0.2 Ma (SHRIMP; Bowring *et al.*, 1993) obtained for the Nemakit-Daldynian deposits of the Olenek Uplift (zircons from basal part of the Kessyusa Formation). The date was originally accepted for the age of the Nemakit-Daldynian Stage lower boundary, but the basal part of the Kessyusa Formation yields fossils characteristic of the stage upper zone, whereas the lower zone corresponds to upper 30-m-thick interval of the underlying Turkut Formation (Khomentovskii and Karlova, 1994b, 2002). Thus, the minimum age limit of the Aim Formation is strictly defined by the dates of 543.9 ± 0.2 and 553 ± 23 Ma. Basaltic flows and sills in boundary strata between the Lakhanda and Ui groups of the Upper Riphean are the youngest objects below the Yudoma Group, which have been reliably dated. The U–Pb dates for baddeleyite from these volcanics are 1005 ± 4 and 974 ± 7 Ma (Rainbird *et al.*, 1998), and their Sm–Nd isochron age corresponds to 948 ± 18 Ma (Pavlov *et al.*, 1992).

Structure of the studied sections. We studied two sections of the Yudoma Group in the northeastern Yudoma–Maya belt, one of the Suordakh Creek, where it flows into the Belaya River, and the other one of the Sakhara River near the Pryamoi Creek mouth (Fig. 1). Samples have been collected from the Lower Yudomian (Yukanda) interval of the Sakhara and Suordakh sections and partly from Upper Yudomian beds of the Suordakh section, which are included here into the Sytyga Formation (Mezentsev *et al.*, 1978). The Lower Yudoma succession of Suordakh section is one of the thickest in the region. It is divided into four local subdivisions: the Nachara, Yalan, Mal, and Tokur formations (Starnikov and Mekhonoshin, 1981; Starnikov *et al.*, 1983), which are considered in this work as subformations of the Yukanda Formation.

The Nachara Subformation (165 m) unconformably overlying the Lakhanda and Ui groups of the Upper Riphean is composed of inequigranular variegate feldspar-quartz sandstones with gravelstone and siltstone interlayers.

The Yalan Subformation (155 m) is composed in its basal interval (18 m) of dark limestone beds with horizontal or wavy lamination, which bear a variable amount of fine siliciclastic admixture. Higher in the section, black calcareous shale and sandstone beds are intercalated with more frequent strata of greenish gray limestones with fine horizontal lamination, which enclose interlayers of cross-laminated oncolitic limestones (60 to 65 m). The overlying gray massive and fine-laminated limestones with rare intercalations of cross-laminated oncolitic, sandy, and stromatolitic varieties (35 m) grade upward into gray to cream-colored micritic and clastic limestones (42 m), which enclose interlayers of sandy dolostone in the lower 30-m-thick interval. In two upper members, there are cherty lenses.

The Mal Subformation (45 m) shows prevalence of variegate horizontal- to cross-laminated limestones, their dolomitic varieties, and dolostones. In these rocks, there are rare interlayers of siltstones and shales, cherty nodules, siliciclastic admixture, finely dispersed hematite, and disseminated pyrite grains sometimes.

The Tokur Subformation (130–135 m) is composed in its lower interval (55–57 m) of light-colored massive dolostones with interlayer of sandy cross-laminated varieties. In the upper interval (75–80 m), it is represented by horizontal- and cross-laminated dolostones, sandy dolostones with desiccation cracks and ripple marks, and by rare flakestones.

In the Sakhara section of Yukanda Formation, which is located closer to the arch of the Early Yudomian dome (Davydov and Sobolev, 1990), the basal terrigenous member is largely reduced in distinction from the Suordakh section and carbonate succession of a lower thickness is lacking variegate rocks, being composed predominantly of dolostones. Basal quartz sandstones and gravelstones of the formation are only 2 to 5 m thick and discordantly overlie here the Upper Riphean Ui Group. These rocks grade upward into succession of five conformable carbonate members, the upper of which is partially eroded and overlain by the Upper Yudomian sandstones. The first member (40 m) is composed of dark gray fine-laminated dolostones with cherty lenses and interlayers of black limestones, the latter getting more frequent closer to the member top. Sandy admixture characteristic of the rocks is most abundant in the lower 10 meters. The second member (60 m) is represented by dark gray horizontal- or cross-laminated silty and dolomitic limestones with oncolitic interlayers in the lower portion and by pale to gray sandy dolostones with sporadic cross-lamination, disseminated pyrite, and cherty lenses in the upper portion. In the third, poorly exposed member (33 m), gray wavy and planar laminated clayey dolostones and limestones reveal presence of silty varieties and cherty lenses. Light gray massive sandy dolostones of the fourth member (50 m) are intercalated with grainstone and flakestone interlayers within the upper part. Flakestones prevail in the fifth member (40 m) of largely cross-laminated sediments with appearing indications of their subaerial exposition. According to composition and C-isotope characteristics (see below), two lower carbonate members of the Sakhara section correspond to the Yalan Subformation of the Suordakh sections, while the third and two upper members are correlative with the Mal and Tokur subformations, respectively (Fig. 1). In both sections, the Yukanda succession recorded a distinct trend of shoaling, i.e., transition from relatively calm lower subtidal settings first to upper subtidal zone with active hydrodynamics, and then, in some terminal horizons, to intertidal environments.

In the Suordakh section, various horizons of the Tokur Subformation are discordantly overlain by the Upper Yudomian terrigenous–carbonate succession (Sytyga Formation) of dominant dolostone members different in thickness (10 to 60 m), which are intercalated with quartz sandstone and siltstone members (12 to 30 m) pinching out in lateral direction. The Cambrian lower boundary is tentatively placed (Mezentsev *et al.*, 1978; Starnikov *et al.*, 1983) at the level of 160–170 m above the succession base. In the northeast of the Yudoma–Maya belt, this boundary is reliably established solely nearby the Sakhara section, where the erosion surface of the Upper Yudomian Sardana Formation is overlain by the variegate carbonate succession that yields near the base the lower Tommotian SSF (Semikhatov *et al.*, 1970; Mezentsev *et al.*, 1978).

INVESTIGATION PROCEDURE

Carbonate rock samples collected during fieldwork to determine C- and O-isotope composition were lacking visual signs of secondary recrystallization. After sawing them in halves, we used one portion for petrographic and cathodoluminescence investigation and the other one for chemical and isotopic analysis. Major ele-

ments of carbonates were analyzed by wet chemical method, and the atomic absorption spectroscopy was applied to determine the Mn and Fe concentrations in the samples dissolved in 1N HCl under room temperature. The Sr concentrations measured by either XRF (using device VRA-10) or isotope dilution method (mixed ^{87}Rb–^{84}Sr as a spike) appeared to be consistent (divergence of results from 1 to 8%).

C- and O-isotope compositions of sampled carbonates have been analyzed mostly at the Harvard University of the United States, and a part of whole rock samples of the least recrystallized (dolo)micrites and (dolo)microsparites has been investigated at the Institute of Precambrian Geology and Geochronology of Russian Academy of Sciences (IPGG). In the last case, CO_2 was extracted from powdered sample aliquots (20 mg in weight) by dissolution of limestones and dolostones in concentrated phosphoric acid (ρ = 1.89 g/ml) under room temperature during 2 and 72 hours, respectively. Mass spectrometer MI-1201 and IPGG standard calibrated with respect to international PDB standards KH-2 and TKL were used to determine C- and O-isotope compositions with analytical uncertainty ±0.2‰ (2σ). At the Harvard University, isotopic analysis was performed for microaliquots separated by miniboring device from those (dolo)micrite segregations of samples which are non-luminescent. The rock powder bored out was dissolved then in concentrated H_3PO_4 (ρ > 1.89 g/ml; t = 90°C), and isotopic composition of extracted and vacuum-distilled CO_2 was measured using three-collector mass spectrometer VG Isogas Prizm. Analytical errors were ±0.2‰ and ±0.4‰ (2σ) for C and O_2, respectively. All $\delta^{13}C$ and $\delta^{18}O$ values are normalized relative to PDB standard.

It is appropriate to mention now how correct is joint application of isotopic data obtained for whole rock samples and microaliquots bored out of them. The results known for Riphean and Vendian carbonates from Namibia, Spitsbergen, and Central Siberia (Fairchild and Spiro, 1987; Knoll and Sweet, 1990; Kaufman *et al.*, 1991; Knoll *et al.*, 1995a; Bartley *et al.*, 2001) show that $\delta^{13}C$ values obtained by two methods are consistent in those cases, when the whole rock samples studied were satisfying geochemical criteria of intact C-isotope systems. For instance, difference between $\delta^{13}C$ values measured using two approaches in proper carbonate rocks from Namibia is 0.3‰ only. In the case of Riphean carbonates from Spitsbergen, it is not greater than 0.5‰ for 70% of studied samples, being as high as 0.5–1‰ for the others. Difference between $\delta^{18}O$ values is therewith 1‰ in average and up to 3–5‰ in some sample pairs.

Criteria used to select carbonate rocks with intact C-isotope systems are the chosen limiting Mn/Sr, Fe/Sr, and $\delta^{18}O$ values and covariation between these geochemical and isotopic parameters (Veizer, 1983; Fairchild *et al.*, 1990; Kaufman *et al.*, 1993; Kaufman, Knoll, 1995, Knoll *et al.*, 1995a; Sochava *et al.*, 1996; Podkovyrov *et al.*, 1998). These criteria are selected because meteoric and elision fluids, the principal agents of C-isotope system violation in carbonates, are enriched in Mn and Fe leached out of siliciclastic rocks and depleted simultaneously in Sr and ^{18}O as compared to seawater. The impact of these fluids usually increases Mn/Sr and Fe/Sr ratios but decreases $\delta^{18}O$ values in limestones and dolostones. Critical Mn/Sr and Fe/Sr ratios, which discriminate samples with disturbed and intact C-isotope systems, are variably assessed in different works. In this work, we accept, like before (Podkovyrov *et al.*, 1998), Mn/Sr ≤ 4 and Fe/Sr ≤ 10 for limestones, and Mn/Sr ≤ 6 and Fe/Sr ≤ 15 for dolostones. The higher values are accepted for the latter, because Sr enters easier the crystalline lattice of calcite than that of dolomite (Veizer, 1983) and Mn represents a trace element captured during the early dolomitization of Precambrian sediments (Brand and Veizer, 1980; Fairchild *et al.*, 1990). As is generally accepted, critical $\delta^{18}O$ value is ≥ –10‰ for Proterozoic carbonates. In early diagenetic dolostones, this value is by 1–3‰ higher than in limestones owing to fractionation of oxygen isotopes in the course of dolomitization, when calcareous sediment was still permeable for diagenetic fluids (Veizer and Hoefs, 1976; Kaufman and Knoll, 1995).

ISOTOPIC–GEOCHEMICAL CHARACTERIZATION OF SAMPLES

In this section, we consider geochemical and other criteria in order to figure out the least altered ("best") samples of the studied collection, which presumably retain C-isotope signature of sedimentation medium.

Yalan Subformation and stratigraphic equivalents. Samples of the Yalan Subformation from the Suordakh section mostly represent limestones typical or slightly dolomitic (Mg/Ca = 0.02–0.25), which are non-luminescent as a rule. Content of siliciclastic admixture in the rocks ranges in general from 1.4 to 7.2% and is much higher in three samples from lower beds (24.4–33.7%) and in one sample from the upper interval (14.6%; see Table 1). The only dolostone sample from the middle part (Mg/Ca = 0.60) bears 24.7% of siliciclastic material. According to geochemical parameters, the subformation is divisible in two rock successions. In the lower one (80–85 m), limestones and their dolomitic varieties are enriched in Mn (235–785 ppm) and especially in Fe (605–13630 ppm), and concentrations of both elements decrease upward in the succession. The Sr concentration ranging usually between 140 and 825 ppm, is extremely high (1650 ppm) in one sample of oncolitic limestone, primarily aragonitic as we think. In the upper succession (50–55 m), the Sr, Mn, and Fe concentrations are much lower: 120–375, 70–205, and 250–1100 ppm, respectively. Ranges of Mn/Sr ratios in limestones of the lower (0.48–2.3) and upper (0.18–1.3) successions are below the critical value that separates the "best" and

Table 1. Geochemical characteristics and isotopic composition of carbon and oxygen in carbonates of the Yudoma Group, the Suordakh section

Sample no.	Thickness, m	Siliciclastic fraction, %*	Mn, ppm	Fe, ppm	Sr, ppm	Mg/Ca	Mn/Sr	Fe/Sr	$\delta^{13}C$, ‰ PDB	$\delta^{18}O$, ‰ PDB
Sytyga Formation										
105-36a	475	2.2	410	6030	112	0.59	3.6	54	0.5	–3.5
D-47	465	1.0	225	4960	120	0.61	1.9	41	1.0	–3.6
105-33	455	27.0	595	16000	130	0.56	4.6	123	–0.2	–7.3
105-32	446	4.6	390	9830	87	0.58	4.5	113	–1.1	–6.0
D-46	440	15.8	730	30400	115	0.61	6.4	265	–0.3	–7.2
D-44	411	6.9	640	11970	95	0.62	6.7	126	3.0	–5.3
105-26	404	3.2	848	10320	250	0.44	3.4	42	3.3	–8.0
105-25	383	7.1	1070	11025	90	0.55	11.8	123	3.5	–8.2
D-39	373	2.5	690	7990	85	0.59	8.1	94	–0.7	–5.2
105-22	368	8.0	355	3890	60	0.58	5.9	65	–4.4	–7.3
105-21	365	2.3	285	2760	65	0.50	4.3	42	3.2	–9.1
Tokur Formation										
D-37	332	8.2	80	2445	60	0.53	1.3	41	–6.8	–6.5
105-19	328	18.5	180	2490	70	0.58	2.5	36	–6.9	–7.7
105-18**	314	7.2	160	1180	112	0.58	1.4	11	–6.8	–8.3
D-34	306	3.2	70	1695	70	0.61	1.0	24	–7.6	–9.9
105-17**	255	7.0	140	1180	125	0.58	1.1	9.4	–6.6	–7.1
D-29	238	16.2	175	3255	120	0.62	1.4	27	–7.8	–7.5
105-16**	231	2.3	90	830	90	0.58	1.0	9.2	–7.8	–7.9
D-27	227	35.4	101	2950	190	0.62	0.55	16	–8.3	–7.3
105-14	212	14.0	160	1825	110	0.53	1.4	17	–7.6	–8.6
105-13**	203	3.1	145	1855	150	0.58	0.96	12	–7.3	–8.6
Mal Subformation										
105-12	195	19.8	215	2515	205	0.40	1.1	12	4.4	–6.9
105-11	182	24.3	235	4100	135	0.59	1.8	30	4.0	–8.2
105-10	176	6.7	170	1280	245	0.17	0.69	5.2	1.9	–8.7
D-12	174	14.1	710	6320	1585	0.11	0.45	4.0	0.8	–9.4
105-9	165	13.5	95	330	215	0.10	0.43	1.5	6.4	–8.0
105-8	163	13.9	192	2050	260	0.28	0.74	7.9	3.2	–9.0
105-7	158	30.9	430	7070	155	0.65	2.8	46	1.2	–8.8
Yalan Subformation										
D-9	150	5.8	70	1100	375	0.11	0.18	2.9	4.9	–9.6
105-4	146	1.4	122	250	190	0.07	0.64	1.3	8.4	–8.8
51-1	144	6.7	150	510	135	0.16	1.1	3.8	5.7	–8.4
51-1	144	6.7	205	725	175	0.16	1.2	4.2	6.0	–9.6
51-2	143	14.6	205	735	195	0.17	1.1	3.8	6.3	–9.3
105-1	139	2.2	145	315	130	0.09	1.1	2.5	5.7	–9.4
*51-3***	137	2.5	90	265	205	0.09	0.4	1.3	6.2	–9.6
*51-4***	131	2.8	155	440	120	0.08	1.3	3.7	8.4	–9.0
*51-5***	122	5.7	150	360	215	0.09	0.7	1.7	6.4	–9.1
51-6	115	24.7	410	435	290	0.60	1.4	1.5	3.5	–7.3
51-7	108	3.0	–	–	–	0.09	–	–	7.2	–9.2
51-7	108	3.0	195	400	170	0.09	1.1	2.3	8.8	–8.9
51-8	106	3.0	140	665	120	0.13	1.2	5.6	7.5	–9.7
D57A	101	7.2	255	1040	180	0.02	1.4	5.8	5.7	–8.9
D-55	80	24.4	260	1020	160	0.02	1.6	6.4	4.7	–8.2
51-12	74	2.6	235	605	355	0.05	0.66	1.7	6.8	–8.7
106-37	66	27.5	330	5610	140	0.18	2.3	40	3.5	–10.7
*51-14***	50	3.0	785	3500	1650	0.02	0.48	2.1	4.1	–9.2
D-54A	45	4.2	450	3860	530	0.12	0.84	7.3	5.0	–9.3
106-36a	20	1.9	410	6010	335	0.10	1.2	18	4.4	–9.3
51-15	18	33.7	670	13630	825	0.25	0.81	17	5.1	–10.0
106-35	11	2.4	460	6550	430	0.11	1.1	15	4.8	–9.5

Note: Italicized numbers denote whole-rock samples subjected to analysis of C- and O-isotope compositions; *siliciclastic fraction insoluble in 1N HCl; **Sr concentration determined by isotope dilution method.

Table 2. Geochemical characteristics and isotopic composition of carbon and oxygen in carbonates from the lower part of the Yudoma Group (Yukanda Formation), the Sakhara section

Sample no.	Thickness, m	Siliciclastic fraction, %*	Mn, ppm	Fe, ppm	Sr, ppm	Mg/Ca	Mn/Sr	Fe/Sr	$\delta^{13}C$, ‰ PDB	$\delta^{18}O$, ‰ PDB
Tokur Formation										
91-32	218	2.6	575	1435	35	0.61	16	41	–3.0	–4.7
91-30	210	2.9	515	2060	55	0.55	9.3	37	–4.2	–6.6
91-29	190	3.6	510	3730	50	0.59	10	75	–3.5	–3.7
91-28	185	15.0	485	1840	30	0.64	16	61	–3.2	–2.5
91-27	175	3.6	480	3520	40	0.64	12	88	–5.4	–5.5
91-26	170	2.8	585	1540	35	0.63	17	44	–5.6	–2.5
91-25	162	0.9	365	1815	30	0.62	12	61	–5.1	–2.7
91-24	155	6.1	435	3290	50	0.58	8.7	66	–6.5	–3.0
91-23	148	3.4	390	2585	50	0.63	7.8	52	–6.6	–3.2
91-22	136	6.4	445	3405	50	0.63	8.9	68	–6.7	–5.8
91-21	135	21.0	170	1575	200	0.59	0.85	7.9	–5.6	–5.3
Mal Subformation										
91-19	125	1.3	415	2230	50	0.25	8.3	44	1.7	–12.6
91-18	121	–	370	3100	60	0.29	6.2	52	1.4	–11.0
91-17	120	6.2	200	3505	95	0.55	2.1	37	1.4	–12.0
91-16	112	3.3	105	825	60	0.61	1.8	14	–1.2	–3.9
Yalan Subformation										
91-15	95	–	130	1200	65	0.61	2.0	19	3.4	–5.9
91-13	76	7.2	120	965	60	0.59	2.0	16	2.0	–7.0
91-11	68	20.7	145	1330	150	0.53	0.96	8.8	3.2	–7.4
91-10	66	17.1	165	1290	110	0.52	1.5	12	2.9	–7.4
91-8	65	8.3	70	335	485	0.09	0.15	0.7	6.1	–8.7
91-6**	63	1.8	55	285	600	0.07	0.10	0.5	6.3	–9.1
91-4	48	6.0	45	320	405	0.23	0.12	0.8	4.8	–7.4
91-1	40	–	55	340	175	0.14	0.31	1.9	6.3	–8.2
89-24	39	6.3	520	1040	115	–	4.5	9.1	6.1	–8.0
89-13	35	7.3	1180	1930	75	0.56	16	26	6.4	–8.6
89-22	22	2.1	345	460	95	–	3.7	4.9	6.3	–8.1
89-19	15	7.3	430	1610	60	0.61	7.2	27	4.5	–7.0
89-9	12	1.0	355	355	145	0.07	2.4	2.4	6.3	–9.8
89-16	10	8.9	220	525	50	0.40	4.3	10	6.0	–9.8
89-8	5	14.4	410	1375	65	0.58	6.3	22	3.9	–4.4
89-15	2	16.1	440	880	55	0.61	8.0	16	3.7	–5.7

Notes: * siliciclastic fraction insoluble in 1N HCl; **Sr concentration determined by isotope dilution method.

altered rocks (Fig. 2). In distinction, Fe/Sr ratios are highly variable in the lower succession (1.7 to 40), exceeding the critical value in four samples (nos. 106-35, 106-36a, 51-15, and 106-37), and more uniform in the upper one (1.3–5.6; see Fig. 3). The mentioned dolostone sample (no. 51-6) from the upper succession is comparatively enriched in Mn (410 ppm) and Sr (290 ppm), being depleted in Fe (435 ppm), and geochemical ratios characterizing the rock are low (Mn/Sr = 1.4; Fe/Sr = 1.5). O-isotope parameters of the rocks are as follows: $\delta^{18}O$ value is –7.3‰ in dolostone and ranges from –8.2 to –9.7‰ in limestones, being lowered to –10.7 and –10.0‰ in two lower Yalan carbonates (samples 51-15 and 106-37), which show a high content of siliciclastic material and high Fe/Sr ratios. Thus, the least altered rocks are represented by four limestone

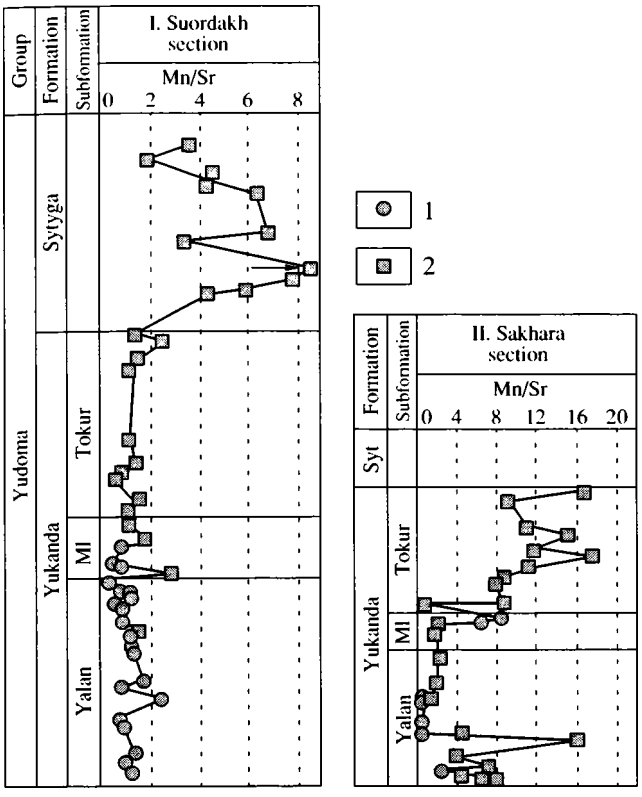

Fig. 2. Variations of Mn/Sr ratio in the Yudoma Group carbonates, Suordakh (I) and Sakhara (II) sections: (1) limestones; (2) dolostones; (Ml) Mal Subformation; (Syt) Sytyga Formation.

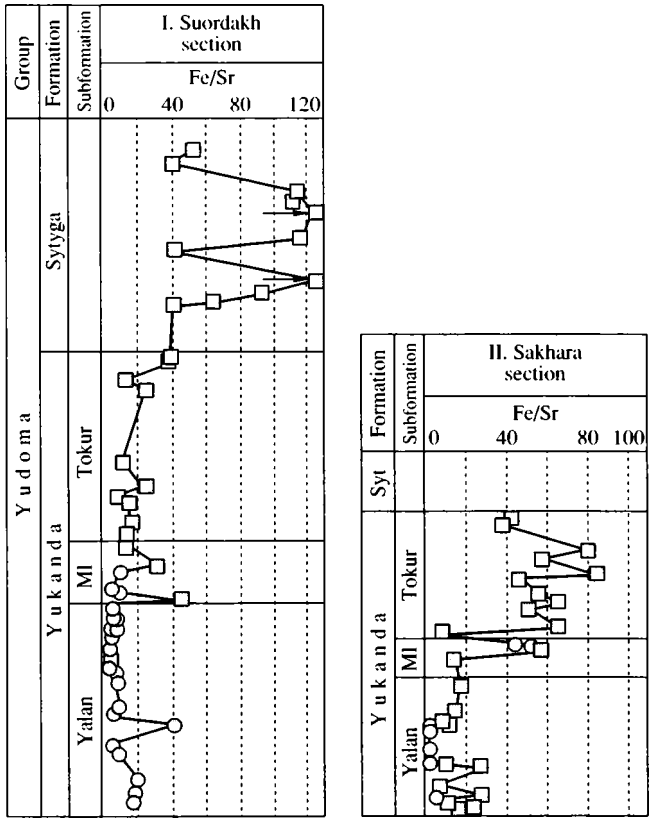

Fig. 3. Variations of Fe/Sr ratio in the Yudoma Group carbonates, Suordakh (I) and Sakhara (II) sections (symbols and abbreviations as in Fig. 2).

samples from basal and middle intervals of the Yalan Subformation in the Suordakh section (nos. 106-35, 106-36a, 51-15, and 106-37).

In the Sakhara section, stratigraphic equivalents of the Yalan Subformation are composed predominantly of typical and calcareous dolostones (Mg/Ca = 0.40–0.61) containing from 6.0 to 20.7% of siliciclastic material (Table 2). Content of this material in normal and dolomitic limestones (Mg/Ca = 0.07–0.23) concentrated here in the subformation middle part is lower in average (1.0–6.0% and 8.3% in one sample). In distinction from dolostones, they show lower concentrations of Mn (45–355 versus 120–1180 ppm) and Fe (285–355 versus 525–1930 ppm) but a higher content of Sr (175–600 versus 50–150 ppm). Like in the Suordakh section, the rocks in question according geochemical parameters are divisible in two successions. In the lower one (39 m), Mn and Fe concentration ranges in dolostones are 220–1180 and 460–1930 ppm, respectively, and one limestone sample contains 355 ppm Mn and 360 ppm Fe. In the upper succession (58 m), we determined 120 to 165 ppm Mn and 965 to 1330 ppm Fe in dolostones versus 45–70 ppm Mn and 285–340 ppm Fe in limestones. Mn/Sr and Fe/Sr ratios are lower in the upper succession than in the lower one (Figs. 2, 3), where both ratios exceed the critical values in four samples (nos. 89-15, 89-8, 89-19, 89-13), three of which have been collected from the basal part. In the upper succession, only Fe/Sr ratios in two dolostone samples (nos. 91-13 and 91-15) are above the critical level, and this may be connected with the high content of siliciclastic material in the rocks. $\delta^{18}O$ values in limestones (–7.4 to –9.9‰) are lower than in dolostones (-4.4 to –9.8‰) being below the critical level in both cases. Thus, in both studied sections the lower Yalan horizons are altered stronger than the upper ones.

Dolostones of the Yalan Subformation were formed at the early diagenetic stage as it follows from isotopic and geochemical characteristics: they are enriched in Sr and show higher $\delta^{18}O$ as compared to those of limestones, although Mn and Fe concentrations in both rock types are comparable. The positive correlation between Mn/Sr and Fe/Sr ratios, on the one hand, and $\delta^{18}O$ values, on the other, which is observable by transition from limestones to dolostones (Figs. 4a, 4b), points to the early diagenetic type of dolomitization as well. At the state of subsidence, the basal Yalan carbonates were partially recrystallized and became enriched in Mn and Fe leached from their siliciclastic fraction and, in the case of Suordakh section, also from underlying sandy–clayey deposits of the Nachar Subformation. Since thick siliciclastic strata of the Lower Yudomian, the

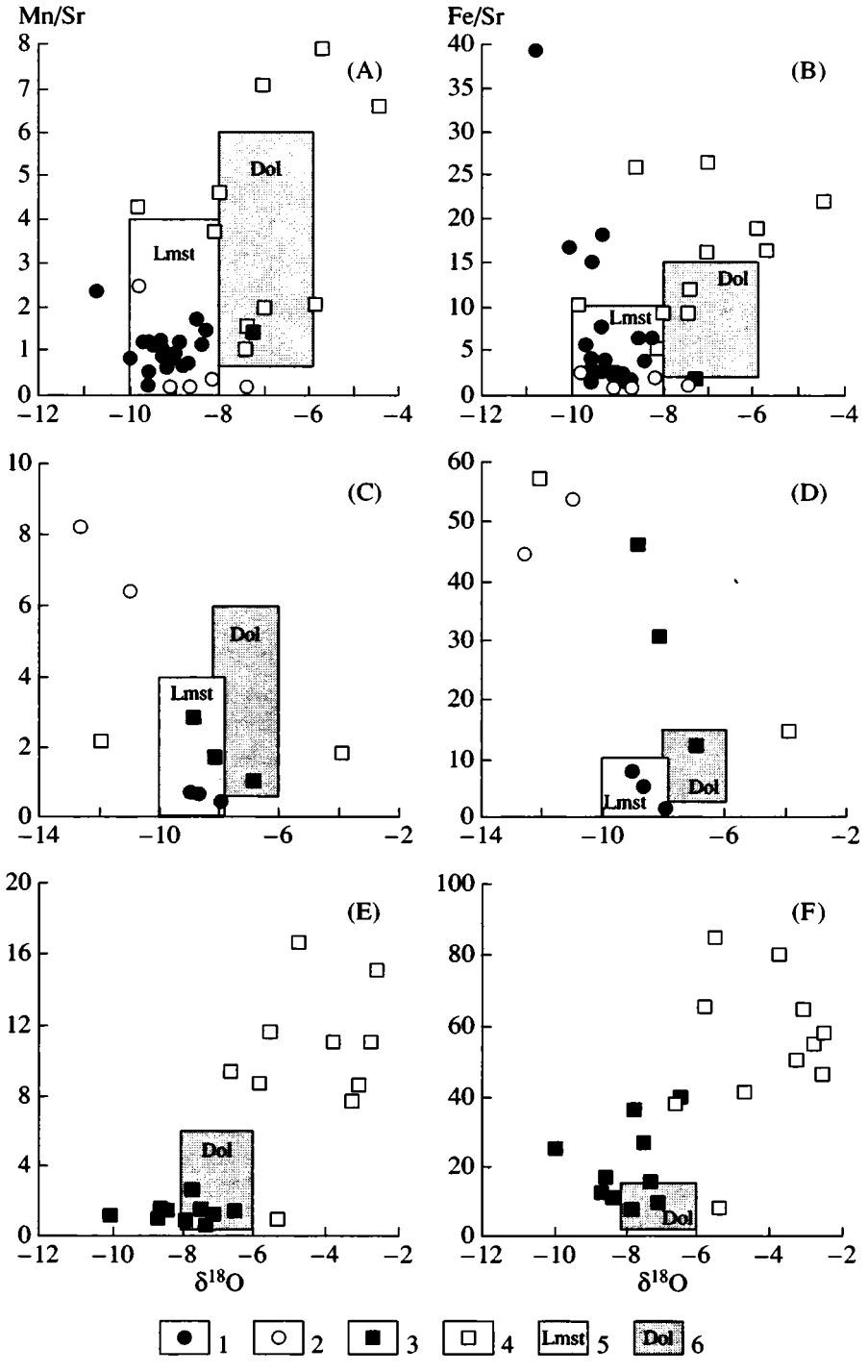

Fig. 4. Covariation diagrams for $\delta^{18}O$ and geochemical ratios (Mn/Sr, Fe/Sr) in Lower Yudomian carbonates, the Yalan (A, B), Mal (C, D), and Tokur (E, F) subformations: (1) limestones of Suordakh section; (2) limestones of Sakhara section; (3) dolostones of Suordakh section; (4) dolostones of Sakhara section; (5) field of the least altered limestones (Lmst); (6) field of the least altered dolostones (Dol).

potential source rocks of Mn and Fe, are unknown in the Sakhara River basin, this factor and less deep subsidence were likely responsible for lower concentrations of both elements in the lower Yalan succession of this area. In addition, different contents of primary aragonite in sediments could be responsible for higher Sr concentrations in the lower Yalan carbonates of the Suordakh section as compared to concurrent rocks of the Sakhara section.

Mal Subformation and stratigraphic equivalents. In the Suordakh section, the variegate Mal Sub-

formation is composed of dolostones near the base (Mg/Ca = 0.65) and in the upper 15-m-thick interval (Mg/Ca = 0.40–0.59), and of limestones in-between (Mg/Ca = 0.10–0.28). The rocks are practically nonluminescent. Content of siliciclastic admixture enriched in finely dispersed hematite is higher in dolostones (19.8–30.9%), than in limestones (6.7–13.9%) and concentrations of Mn (95–430 ppm) and Fe (330–7070 ppm) are proportional to abundance of terrigenous and dolomitic fractions in the rocks. The Sr concentration changes from 215 to 260 ppm in limestones and from 135 to 205 ppm in dolostones. Ranges of Mn/Sr and Fe/Sr ratios in limestones, the dolomitic varieties included, are 0.43–0.74 and 1.5–7.9, respectively, thus being below the critical values. As for dolostones, low Mn/Sr (1.1) and Fe/Sr (12) ratios are established in one sample of these rocks (no. 105-12), whereas two others (nos. 105-7 and 105-11) showed moderate values of first parameter (1.8 and 2.8) and supercritical level of the second one (30 and 46). The negative correlation of both ratios with $\delta^{18}O$ values (Figs. 4c, 4d) is indicative of dolomitization with a freshwater impact that is consistent with $\delta^{18}O$ values in altered (–8.2 to –8.8‰) and almost intact (–6.9‰) dolostones and in associated limestones (from –8.0 to –9.0‰). Thus, two samples (out of seven studied) from the Mal Subformation (nos. 105-7 and 105-11) do not satisfy criteria indicative of the least altered carbonate rocks.

In the upper well-exposed half, the Mal deposits of Sakhara section begins with dolostones (Mg/Ca = 0.55–0.61) grading upward into dolomitic limestones (Mg/Ca = 0.25–0.29). The rocks colored gray contain 1.3 to 6.2% of siliciclastic admixture. Distribution trend of trace elements is different here from that of the Suordakh section, because concentrations of Mn (105–415 ppm) and Fe (825–3505 ppm) are higher in the rocks with a lower content of dolomitic component. In addition, the Sr concentration is lower than in concurrent carbonates of the Suordakh section: 60–95 ppm in dolostones and 50–60 ppm in dolomitic limestones. As a result, Mn/Sr and Fe/Sr ratios in limestones (6.2–8.3 and 45–52 respectively) are above the critical level, whereas in dolostones, the first parameter is below it (1.8–2.1) but the second one (14–37) exceeds the critical value (Table 2; Figs. 2 and 3). In all the rocks except for one dolostone sample (no. 91-16) from the basal portion of sampled interval, $\delta^{18}O$ values are equal to or less than 11.0‰. Consequently, rocks of the Mal interval in the Sakhara River basin were strongly recrystallized at the epigenetic stage and experienced partial dedolomitization. Sample 91-16 is the only one weakly altered (Mn/Sr = 1.8; Fe/Sr = 14; $\delta^{18}O$ = –3.9‰).

Tokur Subformation and stratigraphic equivalents. In both sections, this interval is composed of dolostones and slightly calcareous dolostones (Mg/Ca = 0.53–0.64) with a mottled cathodoluminescence. In the Suordakh section, content of siliciclastic material in dolostones usually ranges from 2.3 to 8.2% except for some horizons, where this material is more abundant (14.0 to 18.0% in three samples and 35.4% in one more sample). Fe concentrations are at the peak values (2490–3255 ppm) in rocks from these horizons (samples D-27, D-29, and 105-19) and range from 830 to 2445 ppm in the others. The level of Mn concentrations is low in general (70–180 ppm), whereas that of Sr concentrations is relatively high for dolostones, usually corresponding to 110–190 ppm and decreasing to 60–70 ppm near the subformation top only. Ratios Mn/Sr in the studied samples (0.55–2.5) are far below the critical value, while ratios Fe/Sr ranging from 9.4 to 41 commonly increase upward in the section and are below the critical value in four samples only (nos. 105-13, 105-16, 105-17, and 105-18). The variation range of $\delta^{18}O$ is from –6.5 to –9.9‰.

In distinction from concurrent rocks of the Suordakh section, the Tokur dolostones of the Sakhara section are relatively enriched in Mn (170–585 ppm) and Fe (1435–3730 ppm), being simultaneously depleted in Sr (not more than 30–50 ppm in general). The only exception is sample 91-21 from a basal horizon of subformation, which contains 200 ppm Sr and shows the Mn concentration minimum. Geochemical ratios in this sample (Mn/Sr = 0.85; Fe/Sr = 7.9) are much below the critical values, whereas in the others they exceed them, sometimes significantly: Mn/Sr = 7.8–17, Fe/Sr = 37–88 (Table 2; Figs. 2 and 3). $\delta^{18}O$ values ranging from –2.5 to –6.6‰ show a weak positive covariation with Mn/Sr, Fe/Sr (Figs. 4e, 4f) and $\delta^{13}C$. This tendency and Mn/Sr ratios growing upward in the section suggest that rocks could be recrystallized under influence of meteoric waters during the pre-Upper Yudomian break in sedimentation. The assumption is inconsistent however with higher $\delta^{18}O$ and $\delta^{13}C$ values characterizing the Tokur dolostones of the Sakhara River area in distinction from their correlatives in the other section. It is possible to assume therefore that the former accumulated in an isolated evaporite basin, although this is inconsistent with geochemical data and inferable paleogeography of the northeastern Yudoma–Maya belt during the Early Yudomian time (Semikhatov *et al.*, 1970; Davydov and Sobolev, 1990). The Tokur sediments of Sakhara section are depleted in Sr as compared to concurrent rocks of the Suordakh section, whereas we should expect the inverse relation assuming their deposition in evaporite basin. It is likely therefore that the Tokur dolomites in the Sakhara River basin were recrystallized under influence of fluids with high $\delta^{18}O$ and $\delta^{13}C$ parameters as in the case of Cenozoic diagenetic dolostones from islands of Atlantic Ocean (Moore, 1989).

Thus, samples 105-13, 105-16, 105-17, and 105-18 with low Mn/Sr and Fe/Sr ratios definitely characterize the least altered rocks in the Tokur Subformation of Suordakh section. We think however that other dolostones from this unit also retain the primary C-isotope signature, because they all showed low Mn/Sr ratios and similar $\delta^{13}C$ values in the case of closely spaced

samples with different Fe/Sr ratios. In the Sakhara section, only sample 91-21 from the Tokur interval appears to be unaltered, as one can judge from geochemical criteria, whereas Mn/Sr and Fe/Sr ratios in the others are extraordinary high, discriminating them from the rest of Lower Yudomian carbonates studied in this work.

Sytyga Formation represents in the studied interval the Upper Yudomian deposits of the Suordakh section. Dolostones from this interval contain 1.0 to 27.0% of siliciclastic material and strongly enriched in Fe (2760–30400 ppm) and Mn (225–1070 ppm). The Fe concentration is growing parallel to increasing abundance of siliciclastic components (Fig. 5a), and the Mn concentration is in covariation with Mg/Ca ratios. Being within the range of 60–130 ppm, the Sr concentration is as high as 250 ppm only in sample 105-26 of calcareous dolostone (Mg/Ca = 0.44). Mn/Sr ratios range from 1.9 to 11.8 and exceed the critical value in four samples (nos. D-39, 105-25, D-44, and D-46), and Fe/Sr ratios are very high (41–265) in all the samples. Changing from –3.5 to –9.1‰, $\delta^{18}O$ values demonstrate a weak positive covariation with Mn/Sr (Fig. 5b) and Fe/Sr ratios. It is clear that C-isotope systems in the rocks are disturbed, because geochemical characteristics imply redistribution of Fe, Mn, and Sr between terrigenous and carbonate components in the course of subsidence and meteoric catagenesis.

Summarizing data considered above, we should mention several important moments. (1) During subsidence of the Yudoma Group deposits, epigenetic alterations in carbonate rocks developed under conditions of closed and open systems, i.e., because of reactions between carbonate and siliciclastic components of the rocks and under influence of elision fluid influx from associated sandy–clayey deposits. (2) The Lower Yudomian (Yukanda) carbonates, especially those of the Suordakh section, are the least altered. About 70 and 40% of samples from that and Sakhara sections, respectively, mostly from the Yalan and Mal subformation in the first case and from the lower Yalan strata in the second case, satisfy the accepted criteria ensuring that rocks are appropriate for subsequent use in C-isotope chemostratigraphic analysis. (3) Rocks of the Tokur interval were altered more intense than underlying Mal and Yalan deposits in the course of meteoric catagenesis during the post-Tokur exsiccation event. (4) The Upper Yudomian deposits of Suordakh section (Sytyga Formation) are so deeply altered during epigenesis that their C-isotope characteristics are unsuitable for chemostratigraphic considerations.

CARBON ISOTOPE COMPOSITION IN LOWER YUDOMIAN DEPOSITS

On the background of the general in-phase character, variations of $\delta^{13}C$ values in the Suordakh and Sakhara sections are of a somewhat different amplitude at correlative levels.

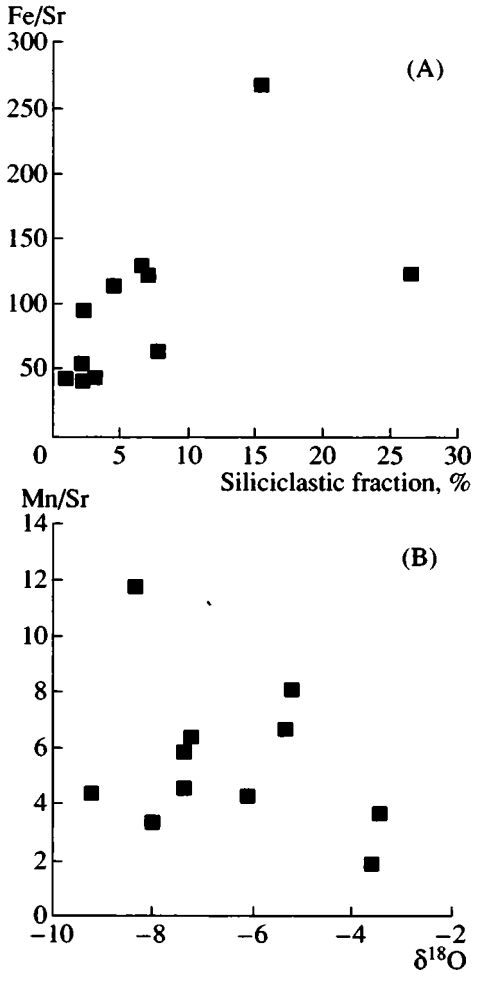

Fig. 5. Covariation diagrams for Fe/ Sr versus percentage of siliciclastic fraction (A) and Mn/Sr versus $\delta^{18}O$ (B) in the Sytyga Formation of Upper Yudomian dolostones.

Yukanda Formation of Suordakh section. The basal 10-m-thick interval of the formation has not been sampled, and three limestone samples from the levels of 11, 18, and 20 m above the base appeared to be altered according geochemical criteria. All samples except for one (no. 106-37), which have been collected higher between 45 and 150 m above the Yalan Subformation base, i.e., within the principal interval composed of normal and slightly dolomitic limestones, are satisfying the accepted geochemical criteria of their retentivity. In these samples, $\delta^{13}C$ values are positive, ranging from +3.5 to +8.8‰ and growing upward in the section. In the middle and upper intervals of the Yalan Subformation, there are four $\delta^{13}C$ maximums (+6.8, +8.8, +8.4 and again +8.4‰), which are separated by relative minimums of positive, though lower values. In distinction from normal limestones, $\delta^{13}C$ values in adjacent dolomitic varieties are by 0.2–0.5‰ lower, but this does nor change the general trend outlined above.

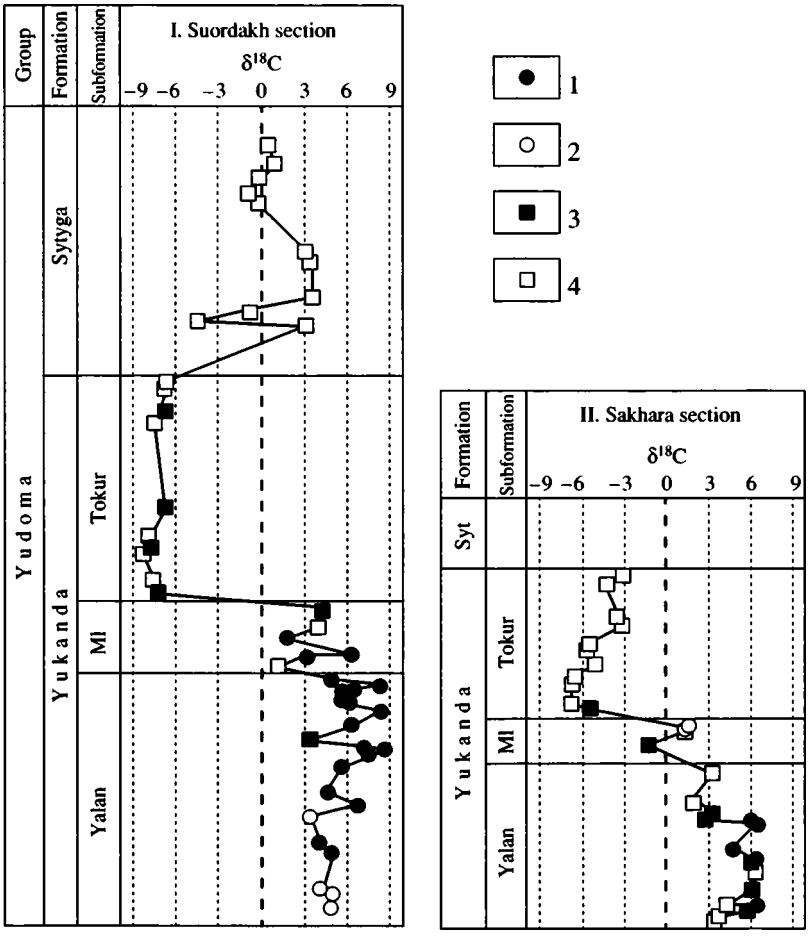

Fig. 6. Secular $\delta^{13}C$ variations in Yudoma Group carbonates of Suordakh (I) and Sakhara (II) sections: samples satisfying (1) and unsatisfying (2) geochemical criteria of the least altered limestones; samples satisfying (3) and unsatisfying (4) geochemical criteria of the least altered dolostones; (Ml) Mal Subformation; (Syt) Sytyga Formation.

Near the Yalan Subformation top, $\delta^{13}C$ drops down to +4.9‰, and further decrease is recorded in the overlying Mal Subformation. In a sample of altered dolostone (Fe/Sr = 46; Mn/Sr = 2.8) that has been collected at the level of 3 m above the Mal Subformation base, $\delta^{13}C$ value is +1.2‰. "Best" samples collected from higher interval of pure and dolomitic limestones depict trend of $\delta^{13}C$ values increasing first up to +3.2 and +6.4‰ and then declining down to +0.8‰ in the middle of subformation and increasing again up to +4.4‰ near its top. Thus within the Mal Subformation, there are recorded two excursions of $\delta^{13}C$ values toward low positive values (Fig. 6). In distinction from Yalan and Mal carbonates, characteristic of which are positive in general though variable $\delta^{13}C$ values, the Tokur dolostones of the Suordakh section demonstrate low negative values of this parameter. In all the samples from this interval, $\delta^{13}C$ values range from –8.3 to –6.6‰, and in the "best" samples they are between –7.8 and –6.8‰ and grow a little toward the subformation top.

Yukanda Formation of the Sakhara section. Two samples of altered dolostones with $\delta^{13}C$ values of +3.7 and +3.9‰, which have been collected 2 and 5 m above the Yalan Subformation base, anticipate values of +6.0 and 6.3‰ in the "best" samples collected 5 and 7 m higher in the section. Consequently, the elevated positive $\delta^{13}C$ values are characteristic of the basal Yalan horizons. In the interval from 10 to 65 m above the subformation base, $\delta^{13}C$ values in limestones and dolostones satisfying geochemical criteria of their retentivity range around +6.2 ± 0.2‰ and are lowered down to +4.8‰ only at the level of 48 m above the subformation base. Two "best" dolostone samples from higher level of 66 and 68 m depict an excursion toward lower values (+2.9 and +3.2‰). In the Suordakh section, the same excursion is recorded in sample 51-6 ($\delta^{13}C$ = +3.5‰) from the correlative stratigraphic horizon. Unaltered rocks, which could characterize the higher 20-m-thick interval of the Sakhara section, have not been found in our collection. Thus, $\delta^{13}C$ values in the Yalan carbonates from this section are lower than in the Suordakh section (+2.9 to +6.3‰ versus +3.5 to +8.8‰) and less fluctuating. This is either a consequence of $\delta^{13}C$ variations within the basin of sedimentation or, what is more

convincing, the impact of decomposed organic matter, because the Yukanda deposits of the Sakhara River basin are enriched in this component.

In the "best" dolostone sample from the middle of Mal Subformation of the Sakhara section, $\delta^{13}C$ is equal to –1.2‰, and in altered rocks of higher horizons this parameter ranges from +1.4 to +1.7‰. As the alteration processes decrease $\delta^{13}C$ values, it is reasonable to suggest that originally characteristic of these horizons were positive values of a moderate magnitude (> 2‰). In such a case, the upper Mal fragment of $\delta^{13}C$ curve characterizing the Sakhara section is comparable in configuration, being displaced toward lower values, with the concurrent Suordakh trend depicted from "best" samples. The negative $\delta^{13}C$ excursion at the middle of the Mal Subformation in the Sakhara section probably corresponds to the upper of two similar excursions established in the Suordakh section (Fig. 6). The Tokur dolostones of the former section are the least altered only in the basal 1.5–2 m thick interval of the unit, and $\delta^{13}C$ value characterizing one sample from this interval is –5.7‰. Accordingly, the boundary between Mal and Tokur subformations marks a transition from low positive to low negative $\delta^{13}C$ values in both studied sections. In the overlaying part of the last subformation represented by altered dolostones in the Sakhara section, these values range from –6.7 to –3.0‰.

Thus, the general trend of in-phase $\delta^{13}C$ variations is characteristic of both studied sections of the Lower Yudomian deposits, and based on this trend the Yukanda Formation can be divided into three parts with different C-isotope signatures. (1) The Yalan carbonates demonstrate relatively high positive though variable (in Suordakh section) $\delta^{13}C$ values and a brief excursion toward lower positive values near the level dividing the subformation into three lower and one upper quarters. (2) The Mal succession of limestones and dolostones, characteristic of which are lower positive $\delta^{13}C$ values and two short-term excursions toward zero and moderate negative values. (3) The Tokur dolostones typical of which are low negative, almost uniform $\delta^{13}C$ values (from –6.8 to –7.8‰). The pre-Upper Yudomian unconformity is the upper limit of the Tokur chemostratigraphic minimum.

Representative C-isotope characteristics of Upper Yudomian deposits are known only for western sections of the four-member Ust-Yudoma Formation nearby the Aldan River (Brasier *et al.*, 1994b, and references therein). In the lower member of that formation, $\delta^{13}C$ values increase upward in the section from –4.5 to –1.5 and –1.0‰. Within the overlying three members of the formation, which belong to the Nemakit-Daldynian Stage, these values ranging between –1.6 and –1.0‰ at the stage base increase up to +3.5‰ near the stage upper boundary and get close to the persistently positive values near the base of *P. antiqua* Zone.

INTERREGIONAL CORRELATION AND STRATIGRAPHIC RANGE OF THE YUDOMIAN

In the last time, C-isotope chemostratigraphy is widely used in correlation of Proterozoic deposits, but this approach can be misleading without assisting control of other chronostratigraphic methods. Possible missteps can be related to the following moments. Fist, fragments of $\delta^{13}C$ variation curve can be of similar configuration at different age levels. Second, criteria used to select the "best" samples for plotting standard variation curves are not uniform, and third, lateral fluctuations of $\delta^{13}C_{carb}$ in upper layer of the World ocean, which are caused by local and regional factors, are possible (Veizer *et al.*, 1999; Melezhik *et al.*, 2001). Chemostratigraphic telecorrelations must be controlled therefore by other methods of chronostratigraphy. At the same time, the composite $\delta^{13}C$ variation curves based on compilation of data, which characterize distant sections, may be potentially misleading, if the reference age levels have not been established in each particular section.

The composite curve of $\delta^{13}C$ variations in the Yudoma Group consists of the lower and upper segments separated by a hiatus, time span of which cannot be figured out from regional data. To evaluate the hiatus range, it is necessary to correlate the Yudoma Group with concurrent deposits in the Olenek Uplift of Siberia and Kalahari craton of Namibia, where their chronostratigraphy is substantiated by a complex of methods.

In the Olenek Uplift, the most complete section of Yudomian deposits is exposed along the Khorbusuonka River (Fig. 7), where they rest unconformably on different Riphean horizons and are divided into four formations, three upper of which yield characteristic fossils (Komar, 1966; Fedonkin, 1987; Knoll *et al.*, 1995a; Khomentovskii and Karlova, 1994b, 2002). In the Maastakh Formation (60 m), the laterally variable basal member of quartz sandstones, gravelstones, and tuffaceous sandstones (up to 20 m) is overlain by the member of microbial, stromatolitic, and less frequent clayey dolostones with cherty nodules (40 m). The Khatyspyt Formation (180 m) is composed of red siliciclastic deposits in the lower part (20 m) and of dark bituminous (up to 1.1% C_{org}) limestones and dolostones intercalated with black shale interlayers in the upper part (160 m). The rocks bear impressions of soft-bodied Metazoa whose assemblages lacking complex ("advanced") morphotypes consist of several forms simple in morphology, which are of wide stratigraphic ranges in Upper Vendian (Fedonkin, 1987, 2003). The Turkut Formation (70 m) of clastic and stromatolitic dolostones and rare limestones bears in the upper interval (30 m) the SSF of the first *A. trisulcatus* Zone of the Nemakit-Daldynian Stage (Khomentovskii and Karlova, 1994b, 2002). In the Kessyusa Formation (50 m) of alternating terrigenous, carbonate, and volcanogenic rocks near the base, the latter bear zircon that is

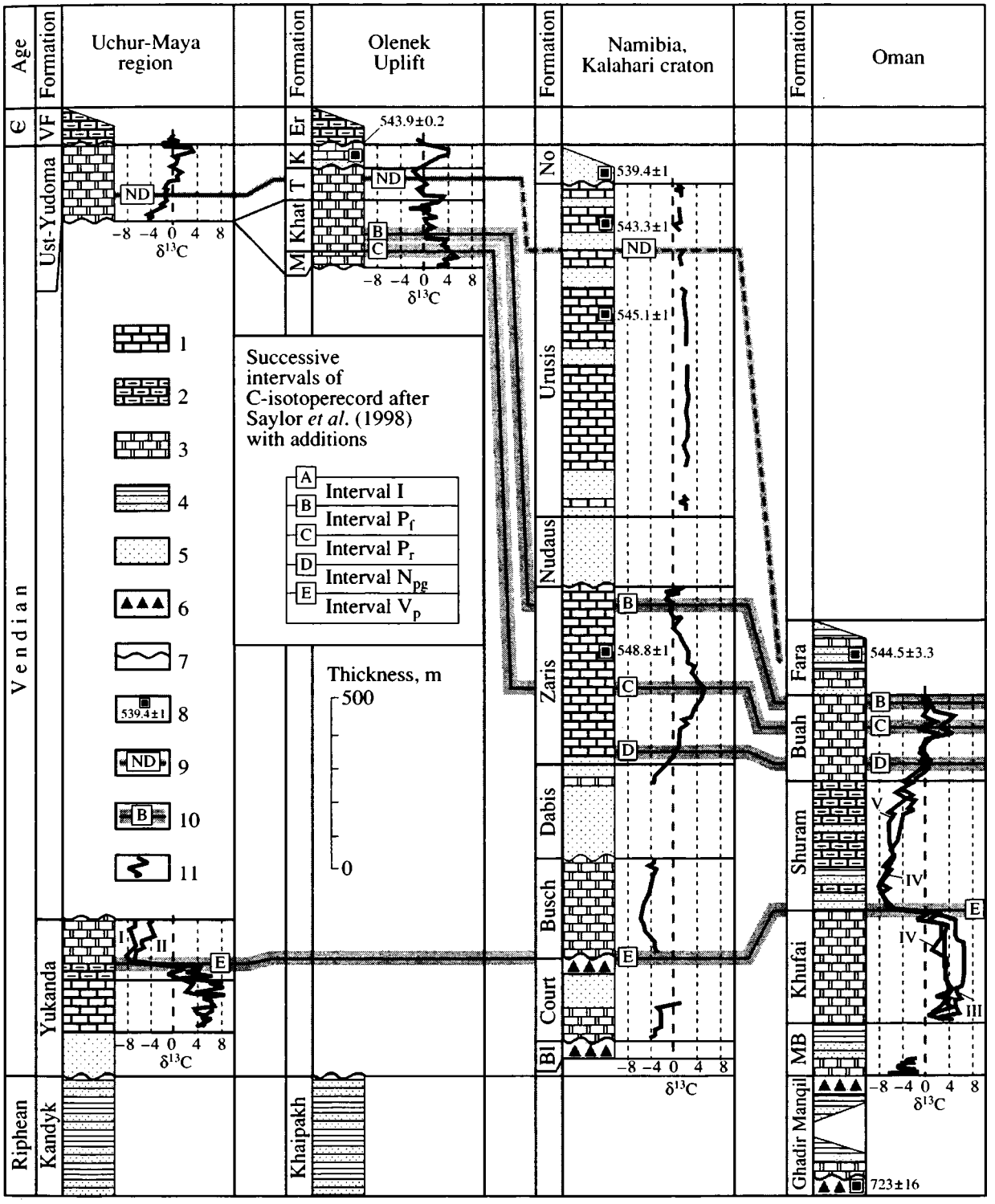

Fig. 7. C-isotope chemostratigraphic correlation and age characteristics of the terminal Precambrian in sections of the Uchur–Maya region (this work and Brasier *et al.*, 1994b), Olenek Uplift (Knoll *et al.*, 1995b), Kalahari craton of Namibia (Kaufman *et al.*, 1991, 1997; Grotzinger *et al.*, 1995; Saylor *et al.*, 1998), and Oman (Burns and Matter, 1993; Burns *et al.*, 1994; Brasier *et al.*, 2000; Amthor *et al.*, 2002): (1) limestones; (2) clayey limestones; (3) dolostones; (4) siltstones and shales; (5) sandstones; (6) tillites; (7) unconformities; (8) U–Pb zircon dates, Ma; (9) lower boundary of the Nemakit-Daldynian Stage; (10) indices of boundaries between intervals of the Vendian C-isotope record below the Nemakit-Daldynian Stage, succession of which is shown in the inset chart (after Saylor *et al.*, 1998, with additions); (11) $\delta^{13}C$ variation curves. Latin numbers denote curves characterizing the following section: (I) Suordakh; (II) Sakhara; (III) Oman Mountains; (IV) Borehole "MQ"; (V) Huqf site. Abbreviations: (Bl) Blaubeker tillites; (Busch) Buschmannsklippe Formation; (Err) Erkeket Formation; (K) Kessyusa Formation; (M) Maastakh Formation; (MB) Masirah Bay Formation; (No) Nomtsas Formation; (VF) Variegate Formation; (T) Turkut Formation; (Khat) Khatyspyt Formation.

543.9 ± 0.2 Ma old according to results of SHRIMP dating (Bowring *et al.*, 1993). In a lower greater part of the formation, there are the SSF of the upper *P. antiqua* Zone of the Nemakit-Daldynian Stage, and 10 m below the top appears fauna of the first (*N. sunnaginicus*) and next of the second (*D. regularis*) zones of the Lower Cambrian Tommotian Stage (Khomentovskii and Karlova, 2002). The Kessyusa Formation bounded by karstified erosion surfaces underneath and on top overlies different horizons of the Turkut Formation (Knoll *et al.*, 1995a; Pelechaty *et al.*, 1996a, 1996b). In the Khorbusuonka River section, $\delta^{13}C$ values were measured in samples, only a half of which satisfies not very strict criteria of their retentivity: Mn/Sr < 8, $\delta^{18}O > -8‰$, $\Delta\delta = (\delta^{13}C_{carb} - \delta^{13}C_{org}) > 30$ (Knoll *et al.*, 1995a). The $\Delta\delta$ values characterize intensity of organic diagenesis, i.e., the early diagenetic oxidation of organic matter and coprecipitation of carbonate carbon enriched in ^{12}C.

In the Maastakh Formation, "best" samples were found in the lower 23-m-thick interval of its carbonate member. Within this interval, $\delta^{13}C$ value first decrease upward from +4.1 to 2.2‰ and then increase up to +4.1 and +3.5‰. Higher in the section, there is 47-m-thick interval composed of altered upper Maastakh dolostones and of basal siliciclastic member of the Khatyspyt Formation. In the upper carbonate succession (140 m) of the last formation, $\delta^{13}C$ values in the least altered samples change in the following manner: they are between +3.2 and +3.6‰ in lower strata (13 m), drop down to +0.3‰ 8 m higher, and are at the level of 0 to +1.5‰ in the next 82 meters of the section except for one sample that yielded value of 2.2‰. In the succession terminal part (38–28 m below the top), there are two negative excursions (–2.6 and –0.9‰) separated by value of +1.0‰, and afterward $\delta^{13}C$ increases up to +2.5‰ at the level of 14 m below the formation top. Knoll *et al.* (1995a) who analyzed $\Delta\delta$ in limestones of the Khatyspyt Formation concluded that these rocks experienced an intense organic diagenesis, and that original $\delta^{13}C_{carb}$ in the rocks were between 0 and +2‰. Samples with low Mn/Sr ratios (0.03–0.23), which depict the aforementioned two negative excursions, could represent an exception. Nevertheless, excursions of this kind are unknown in other sections of the Olenek Uplift (Pelechaty *et al.*, 1996a, 1996b). Moreover, the sample with the lowermost $\delta^{13}C$ value (–2.6‰) is dolomitized (Mg/Ca = 0.101), displaying high $\delta^{18}O$ (–1.6‰), and was likely recrystallized under influence of hydrothermal fluids (Knoll *et al.*, 1995a).

Limestones of the Turkut Formation experienced both the organic diagenesis and the meteoric catagenesis at the time of pre-Kessyusa hiatus, and only 7 of 30 studied samples from this formation satisfy the accepted criteria of their retentivity. In these samples, $\delta^{13}C$ is +3.2‰ at the level of 2 m above the formation base, decreases down to +0.1‰ 1 m higher, and ranges between –1.0 and +0.4‰ in the middle part below the *A. trisulcatus* Zone. In the zone upper part close to the formation top, this parameter in the "best" samples declines (–0.9 to –2.6‰) and then grows again (+2.5 to +4.0‰) in the upper one third of the Kessyusa Formation (terminal part of the *P. antiqua* Zone). C-isotope chemostratigraphy of other Yudomian successions of the Olenek Uplift (Pelechaty *et al.*, 1996a, 1996b) differs from that of the Khorbusuonka section (Knoll *et al.*, 1995a) in some details only, which depend on local peculiarities of epigenetic processes (Pelechaty *et al.*, 1996a) and partially on diverging geochemical criteria used to select the least altered samples from that uplift.

C-isotope characteristics for "best" samples from lower and upper zones of the Nemakit-Daldynian Stage of the Khorbusuonka section are comparable in general with those of the Ust-Yudoma Formation in the Aldan sections (Brasier *et al.*, 1994b). Values of –2.6 and –3.2‰, which have been obtained in sections of the Khorbusuonka (Knoll *et al.*, 1995a) and Olenek (Pelechaty *et al.*, 1996a) for samples from the uppermost part of *A. trisulcatus* Zone, represent an exception. They can be regarded as additional argument in favor of intense secondary alterations in the Turkut carbonates at the time of the pre-Kessyusa hiatus. As is mentioned above, both zones of the Nemakit-Daldynian Stage in two regions are correlated based on paleontological data (Khomentovskii and Karlova, 1992, 1994b, 2002).

In predominantly carbonate 180-m-thick succession of the upper Khatyspyt and lower Turkut formations, which predates the Nemakit-Daldynian Stage, original $\delta^{13}C$ values probably ranged from 0 to +2‰ (Knoll *et al.*, 1995a). In the Uchur–Maya region, considerable intervals of sections with similar C-isotope characteristics are unknown below the Nemakit-Daldynian Stage, i.e., in the first member of the Upper Yudomian Ust-Yudoma Formation (Brasier *et al.*, 1994b) and in the Lower Yudomian Yukanda Formation (this work). Moreover, the Yudoma Group is lacking chemostratigraphic intervals comparable in C-isotope parameters with the Maastakh and basal Khatyspyt strata demonstrating moderate positive $\delta^{13}C$ values. In the Uchur–Maya region, similar values are known in the Yalan Subformation of the Sakhara section only, but this unit cannot be correlated with the Khatyspyt–Maastakh interval, because the composite section includes the extended Tokur interval in between, where $\delta^{13}C$ values are strongly negative, and respective segments of $\delta^{13}C$ variation curves are dissimilar in shape (Fig. 7). Since biostratigraphic correlation of sections under comparison is reliable at the Nemakit-Daldynian level, it is reasonable to assume that the Maastakh, Khatyspyt, and probably the lower part of Turkut Formation of the Olenek Uplift correspond in range altogether to the hiatus between the Aim and Ust-Yudoma formations of the Uchur–Maya region. This conclusion is consistent with data, though fragmentary, which imply that characteristic of the upper part of the Lower Yudomian deposits exposed in the western part of above region is a moder-

ate level of negative $\delta^{13}C$ values from –3.7 to –0.3‰ (Pokrovskii, 1996).

A significant hiatus in the middle of the Yudomian type succession is also evident from its correlation with Vendian deposits of the Kalahari craton in Namibia. These deposits corresponding to the Blaubeker tillites, Witvlei Group, and Kuibis and Schwarzrand subgroups of the Nama Group (Fig. 7) have combined C-isotope, isotopic-geochronological and paleontological characteristics and have been frequently considered in a global chemostratigraphic aspect (Kaufman *et al.*, 1991, 1997; Saylor *et al.*, 1998; Walter *et al.*, 2000; Brasier *et al.*, 2000; and other works). The Blaubeker tillites basal in the succession are discordantly overlain by the Witvlei Group and have been correlated with lower tillites of the Varangerian glaciohorizon. The group succession is compiled based on correlation of two distant sections, one of which includes the terrigenous–carbonate Court Formation (up to 200 m) and predominantly carbonate Buschmannsklippe Formation (up to 320 m) separated from each other by unconformable boundary. In the other section, the Blässkranz tillites (200 m) and overlying carbonate deposits (110 m) are exposed in a series of tectonic slices. The carbonate rocks are correlated based on lithologic and C-isotope characteristics with a lower part of the Buschmannsklippe Formation, whereas the Blässkranz tillites correlated with the upper Varangerian glaciohorizon are placed at the level of hiatus between this and underlying Court Formation (Kaufman *et al.*, 1991; Saylor *et al.*, 1998).

The Court Formation overlying the Blaubeker tillites is composed of layered dolostones in the lower part and of siliciclastic sediments with rare interlayers of stromatolitic dolostones in the upper one. There are presumably the lacustrine and lagoonal facies among the formation deposits (Kaufman *et al.*, 1991, 1997), and carbonates of this subdivision are intensely altered ($\delta^{18}O < -10‰$ and/or $Mn/Sr > 2$; Saylor *et al.*, 1998). In the lower part of the formation, $\delta^{13}C$ value increase upsection from –4.3 to –1.5‰, whereas in the upper part positive values appearing in addition to negative values approach +5.8‰ in a sample with acceptable geochemical parameters. Marine facies prevailing in the formation and a uniform trend of $\delta^{13}C$ variations defined by samples with variable geochemical parameters suggest, as we think following Saylor *et al.* (1998), that carbonates of the Court Formation recorded in general the initial C-isotope composition. These rocks are interpreted as the cap dolomites of the Blaubeker tillites. The Buschmannsklippe Formation (320 m) terminal one in the Witvlei Group is represented by dolostones and limestones with interlayers of terrigenous sediments in the middle interval, where carbonates are fairly altered ($\delta^{18}O < -10‰$ and $Mn/Sr > 2$), while below and above that interval they are better preserved (Saylor *et al.*, 1998). Characteristic of the formation are negative $\delta^{13}C$ values, which decrease from –2.6 to –5.5‰ in the lower portion and then increase toward the top up to –3.5‰.

The Nama Group conformably overlies the Buschmannsklippe Formation in a central part of its distribution area, truncates the Witvlei Group and rests with angular unconformity on the pre-Vendian rocks along periphery of the area. The basal Dabis Formation (300 m) of the group (the first unit of the Kuibis Subgroup) consists mostly of terrigenous rocks, although in the upper part it includes the Mara Member of dolostones (80 m) occurring in places. In "best" samples from this member, $\delta^{13}C$ values range from –4. to –3.5‰ and are therefore similar to those of the uppermost Buschmannsklippe strata. The Mara member yields *Cloudina* shells first occurring at this level, while soft-bodied Metazoa first appear in the overlying Kliphoek Member, the terminal one in the Dabis Formation. Two groups of fossils mentioned above are confined respectively to carbonate and siliciclastic facies alternating in the section. They also occur higher in the section till uppermost horizons of the Urusis Formation, the Schwarzrand Subgroup (Grant, 1990; Narbonne *et al.*, 1997; Saylor *et al.*, 1998; Grotzinger *et al.*, 2000).

An important trait of section in Namibia is transition from uniform negative to growing positive $\delta^{13}C$ values across the conformable boundary between the Dabis and overlying Zaris formations (up to 500 m), the latter composed of limestones with sandstone and shale interlayers. In slightly altered limestones near the Zaris Formation base, $\delta^{13}C$ values range from +0.2 to +1.2‰ and then, in "best" samples from higher stratigraphic levels, they grow to attain maximum somewhat below the formation midpoint (+5 or +6‰ in different sections). The maximum is recorded slightly below a tuff horizon, the U–Pb age of zircons from which is 548.8 ± 1 Ma (Grotzinger *et al.*, 1995; Saylor *et al.*, 1998). In the upper part of Zaris Formation, $\delta^{13}C$ values are reduced to 0 ± 1‰ in "best" and altered samples, the latter collected from terminal horizons. Values close to +2‰ appear in variably altered samples collected near the formation top. They signify a transition to next interval of the section, to the Schwarzrand Subgroup with its specific C-isotope parameters.

The Nudaus Formation (~200 m), the basal one in the Schwarzrand Subgroup, is composed of terrigenous rocks and underlies the thick Urusis Formation (up to 1100 m) that is divided into three successions with sequence boundaries in between (Saylor *et al.*, 1998). These are (1) shales and fine-grained sandstones with rare dolostone interlayers 220 m thick in total; (2) diverse 450-m-thick limestones, stromatolitic varieties included, with a minor shale–siltstone member in the upper part; (3) a succession (about 500 m) of four limestone members 50 to 100 m thick each with three thinner (30 to 70 m) terrigenous members separating them. In the lower part of the first and near the base of the third carbonate members, there are tuff interlayers with zircons, the U–Pb ages of which are 545.1 ± 1 and

543.3 ± 1 Ma respectively (Grotzinger *et al.*, 1995). In "best" carbonate samples from subdivisions (1) and (2), $\delta^{13}C$ values are close to +2‰, whereas in two lower carbonate members of subdivision (3) they range from +0.9 to +1.9‰. In carbonates of the third member, this parameter is close to +1‰ except for one sample with $\delta^{13}C = -0.1$‰. Altered limestones ($\delta^{18}O < -10$‰) at the top of third and in the fourth member showed $\delta^{13}C$ values from 0 to +2‰ (Kaufman *et al.*, 1991, 1997; Saylor *et al.*, 1998).

The Schwarzrand Subgroup is terminated by the Nomtsas Formation of carbonate and siliciclastic sediments, which filled in deep erosional incisions in deposits of subdivisions (2) and (3) of the Urusis Formation. The Nomtsas Formation bears trace fossils *Trichophycus (=Phycodes) pedum*, and volcanogenic zircons from its basal level are 539.4 ± 1 Ma old according to results of U–Pb dating (Grotzinger *et al.*, 1995). Samples used to depict a fragmentary C-isotope characterization of the formation are altered in general ($\delta^{18}O < -11$‰). In the better preserved whole-rock limestone samples ($\delta^{18}O > 9.4$‰) from middle interval, $\delta^{13}C$ values are +2.4 and +1.8‰, and in bored micrite fractions they correspond to +2.05 and +1.90‰.

In the cited works, the Nomtsas Formation is attributed to the Lower Cambrian based on occurrence of *T. pedum*. We doubt this inference because of several reasons. First, the occurrence levels of *T. pedum* in different regions are asynchronous, from the base of the Nemakit-Daldynian Stage of the Vendian to the Atdabanian and Botomian stages of the Lower Cambrian (Rozanov *et al.*, 1997). Second, the U–Pb age of zircons from the Nomtsas Formation base is 539.4 ± 1 Ma, and one can see here an analogy with the upper interval of the Nemakit-Daldynian Stage, what is consistent with $\delta^{13}C$ values in "best" samples from the formation middle part. Third, the U–Pb age of zircons from terminal part of the underlying Urusis Formation (543.3 ± 1 Ma) is identical, within the analytical error, to the age of zircons (543.9 ± 0.2 Ma) from boundary beds of lower and upper zones of the Nemakit-Daldynian Stage in Siberia. In other words, the Nomtsas Formation is correlative, up to its middle level at least, with the upper zone of that stage, and the stage lower boundary is in Namibia at the terminal level of the Urusis Formation a little below the horizon with zircons dated back to 543.3 ± 1 Ma (referred to further as Horizon 543).

With this conclusion at hands, we meet the following problem however. Deposits of the Urusis Formation in proximity to Horizon 543 are devoid of variable $\delta^{13}C$ values characteristic of the Nemakit-Daldynian Stage, which range from –1.6‰ at the base of the stage lower zone to +3.5‰ at the top of the upper one. Facts important in this case are as follows. In "best" carbonate samples from the Urusis Formation interval enclosing Horizon 543, $\delta^{13}C$ values (from –0.1 to +1.3‰) are comparable with those measured in boundary beds of two Nemakit-Daldynian zones in Siberia(–0.3 to +1.4‰; Brasier *et al.*, 1994b). Values from 0 to +2‰ prevailing above the horizon 543 near the Urusis Formation top are measured in carbonates, which are rather strongly altered here (Saylor *et al.*, 1998). It is logical to connect these alterations with meteoric catagenesis during the break in sedimentation before accumulation of the Nomtsas Formation. In addition, value of +2.4‰ determined in the overlying Nomtsas Formation fills to some extent the gap of moderate positive values, which are characteristic of the Nemakit-Daldynian Stage upper part (about +3.5‰) but have not been established near the Urusis Formation top. As for negative $\delta^{13}C$ values typical of the base of *A. trisulcatus* Zone and underlying strata of Ust-Yudoma Formation in Siberia, their absence below Horizon 543 in Namibia can be connected with siliciclastic composition of the correlative horizon in the Urusis Formation. It is possible that this siliciclastic member exactly is the basal one in the Nemakit-Daldynian interval of the section under consideration (Fig. 7).

Unfortunately, the soft-bodied fossils occurring in Namibian and Siberian sections are unsuitable for a detailed stratigraphic correlation because of three reasons. First, fossils of this kind have not been found so far in the Nemakit-Daldynian deposits of Siberia. Second, forms of complex morphology (*Pteridinium, Rangea, Ernietta, Swartpuntia*) characteristic of the Nama Group are unknown in Vendian carbonate deposits of Siberia. Third, *Cyclomedusa* and *Nemiana*, the only genera in common for the Nama Group and Khatyspyt Formation, are of a wide stratigraphic range within the Upper Vendian (Fedonkin, 1987, 2003).

Based on C-isotope characteristics of Namibian successions, which are considered above, Saylor *et al.* (1998) modified chemostratigraphic subdivision of Vendian deposits, which was established in the Olenek Uplift for their upper interval (Pelechaty *et al.*, 1996a, 1996b). The subdivision scheme they suggested includes four chemostratigraphic intervals. The lower interval N_{pg} (Negative postglacial) of rather low negative $\delta^{13}C$ values (down to –5.5‰) is limited at the top by level, where they are close to zero (Buschmannsklippe and Dabis formations). Second interval P_r (Positive rising) is that, where values increase from 0 to +6‰ (lower part of the Zaris Formation). Third interval P_f (Positive falling) is characterized by decrease of $\delta^{13}C$ values from +6‰ down to zero (the greater upper part of the Zaris Formation except for terminal 40 m). Finally the extended upper interval I (Invariant), where negative values of about –1‰ near the Zaris Formation top rapidly turn into positive values of +2 to +1‰, which remain stable higher in the section. Leaving aside the preservation state of rocks and sampling frequency, it is reasonable to attribute to this interval the upper 40 m of the Zaris Formation and the overlying Schwarzrand Subgroup except for its terminal Nomtsas Formation (Saylor *et al.*, 1998). However, considerations presented above imply that interval I *sensu*

stricto should be limited on top by the pre-Nemakit-Daldynian level of that subgroup.

The comparative analysis of $\delta^{13}C$ variations shows that chemostratigraphic intervals I, P_f, and P_r defined in Namibia, equivalents of which are established in a series of distant Vendian successions (Saylor *et al.*, 1998; Walter *et al.*, 2000; Brasier *et al.*, 2000), are also recognizable in sections of the Olenek Uplift (Fig. 7). The global rank of these chemostratigraphic subdivisions leaves no doubts that their absence in the Yudomian succession of the Uchur-Maya region is indicative of a hiatus between the Lower and Upper Yudomian deposits. Isotopic dates known for the upper part of Zaris Formation (548.8 ± 1 Ma, the mid-P_f level) and for the pre-Nemakit-Daldynian strata of the Urusis Formation (545.1 ± 1 Ma) suggest that the above break in sedimentation was 4 to 4.5 m.y. long at least.

In the Olenek Uplift, equivalents of the interval N_{pg} are unknown, but they are reliably established in the Uchur–Maya region in the range of the Tokur Subformation of the Yukanda Formation (Fig. 7). Thus, the Tokur Subformation corresponds to the Buschmannsklippe and Dabis formations of Namibia, where C-isotope chemostratigraphy of underlying deposits, the Varangerian glaciohorizon included, is studied inadequately, thus being inappropriate for telecorrelation. To solve the problem of relation between that glaciohorizon and Yudoma Group, it is necessary to correlate the latter with the Huqf Supergroup of Oman.

The thick (up to 5 km) Huqf Supergroup of siliciclastic and carbonate deposits is crowned by evaporite succession. Sediments of the supergroup discordantly overlie crystalline rocks of basement with magmatic zircons 830–780 Ma old according to results of U–Pb dating and are bounded at the top by unconformity separating them from overlying Upper Cambrian deposits (Gorin *et al.*, 1982; Burns and Matter, 1993; Brasier *et al.*, 2000). Zircons with U–Pb age of 723 ± 16/–10 Ma are from tuff horizons near the supergroup base, while in the terminal part and correlative deposits zircons from similar horizons are 544.5 ± 3.3, 542.6 ± 0.3, and 542.0 ± 0.6 Ma old (Brasier *et al.*, 2000; Amthor *et al.*, 2003).

The Huqf Supergroup is divided from the base upward into the Abu Maharah, Nafun, and Ara groups. The lower portion of the first group corresponds to the Ghadir Manqil Formation (up to 1.8 km) of the two-member structure. The basal Ghubrah Member (~300 m) of tillites with interbeds of volcanic rocks and rare sandstones is crowned by a bituminous limestones and dolostones, which are interpreted as cap carbonates of Ghubrah tillites and show $\delta^{13}C$ values quickly rising upward in the section from –3 to +7‰ (Brasier *et al.*, 2000). The Ghubrah tillites enclosing the tuff horizon that is 723 Ma old belong to the Sturtian glaciation. The upper Fiq Member (up to 1.5 km) of the Ghadir Manqil Formation is presumably bounded by unconformity from below, being composed of alternating packages of tillites (up to 100 m), sandstones and laminated siliciclastic sediments (up to 50 m) with dropstones, and rarer conglomerates, siltstones, and shales (Brasier *et al.*, 2000; Leather *et al.*, 2002). Rb–Sr ages for trachyte and probably comagmatic granitoids occurring in the member range from 562 ± 42 to 556 ± 11 Ma (Burns and Matter, 1993), being rejuvenated as it seems. The K–Ar date of 654 ± 12 Ma (Gorin *et al.*, 1982) obtained for the same volcanic rocks is of ambiguous meaning. The Masirah Bay Formation (80–100 m) terminating the Abu Maharah Group is composed of siltstones and shales with the 8-m-thick Hadash Member of dolostones at the base, in which $\delta^{13}C$ values are –6.5 to –5.0‰ near the lower boundary and –2.3 to –1.5‰ at the top. These dolostones representing sediments of post-glacial transgression are interpreted as typical cap carbonates and overlie, besides the Fiq Member, the older deposits as well (Brasier *et al.*, 2000; Leather *et al.*, 2002). As for the Fiq Member of glacial deposits, it is attributed to the Varangerian (Marinoan) level based on chemostratigraphic parameters of the overlying deposits (Brasier *et al.*, 2000).

The Nafun Group above the Masirah Bay Formation corresponds to conformable succession of the Khufai, Shuram, and Buah formations. The Khufai Formation (up to 400 m) of normal and bituminous dolostones includes anhydrite interlayers in the lower part. The Shuram Formation (up to 750 m) is composed of siltstones with rare interbeds of oncolitic limestones in the lower part and of alternating shales, clayey limestones and flakestones in the terminal interval. Limestones and dolostones of the Buah Formation (up to 600 m) enclose stromatolitic bioherms and shale interlayers in the lower part and contain abundant cherty nodules, pseudomorphs after gypsum crystals, and anhydrite and sandstone interlayers in the upper one (Gorin *et al.*, 1982; Burns and Matter, 1993; Burns *et al.*, 1994).

C-isotope chemostratigraphy of the Nafun Group was studied in natural outcrops in northern (Oman Mountains) and eastern (Huqf site) areas and in two borehole sections drilled on flanks of saliferous fields in north- and southeastern regions of the country (boreholes "MQ" and "T" respectively). The carbonate rocks studied are altered to a variable extent. For instance, in the Oman Mountains $\delta^{18}O$ values in rocks of the Shuram Formation and at the base of Buah Formation range from –14 to –10‰, whereas in Borehole "T" the same parameter usually changes from –8 to –6‰ in the Khufai–Shuram boundary beds and rises upward to be –4 to –2‰ in the Buah Formation. In distinction, $\delta^{18}O$ values characterizing the Huqf section are –6 to –2‰ in the Khufai Formation being as low as –6 to –7‰ in the Shuram Formation, whereas in Borehole "MQ" they range from –8 to –4‰ in the lower interval of the group and from –4‰ to zero in the upper one (Burns and Matter, 1993). Published data on Mn and Sr concentrations (Burns *et al.*, 1994) in a third part of samples used for C-isotope analysis by Burns and Matter (1993) who

studied the Huqf section confirm the variable alteration degree of the rocks. In this section, Mn/Sr ratios range between 0.5 to 27.1 in general, being minimal (0.5–4.1) in the upper Khufai and Shuram formations and maximal (9.4–27.1) in the Buah Formation. Unfortunately, $\delta^{13}C$ variation curves in four sections of the Nafun Group (Burns and Matter, 1993; Brasier *et al.*, 2000) have been plotted without due account for the alteration degree in the rocks that is why the curves are not compatible in a strict sense (see Fig. 3 in Burns and Matter, 1993). Taking this feature and sampling frequency into account, we consider sections of Borehole "MQ" and Huqf site as most representative for the Nafun Group as a whole (Fig. 7) and section of the Oman Mountains as additionally substantiating chemostratigraphy of the Khufai Formation.

Characteristic of the Khufai Formation are rather high $\delta^{13}C$ values, which are at maximum (+5.5 to +7.5‰) in samples with $\delta^{18}O \geq -10$‰ from the Oman Mountains. In the Huqf and "MQ" sections, they range from +3 to +5‰ in samples with $\delta^{18}O > -8$‰. In the Huqf section, the above variation range is typical of samples with low (< 4.1) and high (7.2–11.4) Mn/Sr ratios. The remarkable chemostratigraphic peculiarities of the Khufai Formation are as follows. (1) In its basal interval, $\delta^{13}C$ increases from < +2‰ to rather high positive values. (2) In the upper one third of "MQ" section of the formation this parameter drops from +3‰ down to zero, and then it rises again to be as high as +2.5 to +3‰; a similar interval of $\delta^{13}C$ decline is recognized in the Oman Mountain at the correlative stratigraphic level. (3) In "MQ" and Huqf sections, negative excursions of $\delta^{13}C$ down to –2‰ and zero, respectively, are recorded near the Khufai Formation top, and afterward $\delta^{13}C$ rises up to zero in the "MQ" section and becomes as high as +2.5 to +3‰ in the Huqf section.

An important change of C-isotope parameters in the Oman sections is established near the boundary separating the Khufai and Shuram formations. At this level, moderate positive $\delta^{13}C$ values turn into low negative within the interval of few meters, and the latter are characteristic of the overlying considerable interval of the Shuram Formation. In Borehole "MQ," $\delta^{13}C$ values of –10 to –7‰ are typical of the lower one third of that formation, and near its top they range from –5 to –3‰. In the Huqf section, this parameter is equal to –8.3‰ in "best" limestone samples (Mn/Sr < 1.1, $\delta^{18}O > -6.9$‰) from the lower interval of that formation, ranges from–7 to –6‰ in the upper one and corresponds to –3.9‰ near the top (Burns and Matter, 1993; Burns *et al.*, 1994). Negative $\delta^{13}C$ values of about –2‰ still characteristic of basal level of the overlying Buah Formation quickly grow upward to be at maximum close to the middle of the unit (+4 to +6‰ in "MQ" section and +2 to +2.5 in Huqf section). In the Buah Formation upper part, they decrease and range in the above sections from +1 to 0 and occasionally to –1‰ near the top.

In southern and central areas of Oman, the Huqf Supergroup is terminated by the Ara Group (up to 1700 m) that is composed of carbonate members 10 to 150 thick (from A-0 to A-6) alternating with thicker members of evaporite (halite and/or anhydrite) deposits. Somewhat below the middle of the group, there are tuff horizons near the top of Member A-3 and close to the base of overlying Member A-4, which yielded zircons with U–Pb ages of 542.6 ± 0.3 and 542.0 ± 3 Ma, respectively. In members from A-0 to A-3, $\delta^{13}C$ values usually range from +2.5 to +3.5‰ in several borehole sections declining sometimes down to +1‰ and zero. In Member A-4, they sharply decrease down to –6 or 7‰ near the base, being higher from –2.5 to –2.0‰ near the top, whereas characteristic of Member A-5 is the value range of +2.5 to +3.5‰ (Amthor *et al.*, 2003). The Ara Group is correlative, based on stratigraphic position and fragmentary C-isotope characteristics, to the Fara Formation of the Oman Mountains, where it is composed of black cherts, dolostones, sandstones, and volcaniclastic rocks. Judging from single determinations, $\delta^{13}C$ values are within the range of +2 to +3‰ in the lower interval of this formation, decrease to the level of –2 to –3‰ in the middle part, and increase again up to +3‰ in the upper one. The U–Pb age of zircons from a tuff horizon near the formation middle level is 544.5 ± 3.3 Ma (Brasier *et al.*, 2000). In the middle part of Fara Formation and in members A-1 to A-3 of the Ara Group, there were found *Cloudina* shells (Burns *et al.*, 1994; Brasier *et al.*, 2000; Amthor *et al.*, 2003). This form occurring in the Upper Vendian of several regions does not rise higher in the Tommotian Stage (Conway Morris *et al.*, 1990; Brasier *et al.*, 2000; Zhuravlev *et al.*, in press).

The presented isotopic-geochronological, paleontological, and stratigraphic data strongly suggest that age limits of considered sections in Siberia, Namibia, and Oman are similar, while available data of C-isotope chemostratigraphy elucidate important details of correlation between the sections.

As is shown above, $\delta^{13}C$ variations within the Khufai Formation of Oman are similar in configuration and amplitudes to those established for the Yalan and Mal (abrupt negative excursion) subformations of the Yukanda Formation in the Uchur–Maya region. This similarity in C-isotope signatures of indicated subdivisions in two regions suggest their correlation, and this conclusion is confirmed by the change of predominantly positive to persistently negative $\delta^{13}C$ values, which is recorded across the Khufai–Shuram boundary, on the one hand, and across the Mal–Tokur boundary, on the other. Nevertheless, the $\delta^{13}C$ rising trend at the base of Khufai Formation, which continues trend of this kind recorded below in the Hadash cap dolomites, is unknown at the base of the Yalan Subformation, where $\delta^{13}C$ values are positive and relatively high. It is logical to think therefore that basal Yalan horizons are younger than those of the Khufai Formation, which recorded the post-glacial $\delta^{13}C$ rise in seawater. It is particularly clear

that the basal Yalan carbonates are younger than the lower Varangerian cap dolomites, characteristic of which are low negative, though growing upward $\delta^{13}C$ values. As for the Shuram Formation and lower member of Buah Formation, typical of which are low negative $\delta^{13}C$ values, these subdivisions of the Oman succession appear to be correlative, based on C-isotope parameters, with the Buschmannsklippe and Dabis formations of Namibia (type units of the interval N_{pg} after Saylor *et al.*, 1998) and with the Tokur Subformation of the Yukanda Formation in the Uchur–Maya region. The overlaying main lower part of the Buah Formation of Oman, where $\delta^{13}C$ is growing first and declining afterward, is correlative with intervals P_r and P_f (probably coupled with the lowermost portion of interval I) corresponding in Namibia to the Zaris Formation, its upper member excluded (Saylor *et al.*, 1998), whereas in the Uchur–Maya region the correlative deposits are missed from the succession (Fig. 7). Judging from C-isotope parameters and superposition of deposits, lower carbonate members A-0 to A-3 of the Ara Group, where $\delta^{13}C$ values are uniformly positive and low, can be correlated with interval I, whereas Member A-4, where these values increase from –7 to –2‰ upward in the section, is comparable with the pre-Nemakit-Daldynian interval of the Ust-Yudoma formation with values growing from –4.5 to –1.0‰ (Brasier *et al.*, 1994b). Accepting such a correlation, we meet a problem however. On the one hand, zircons from members A-3 and A-4 are 542.6 ± 0.3 and 542.0 ± 0.3 Ma old and suggest that respective middle interval of the Ara Group is correlative with the upper zone of Nemakit-Daldynian Stage, which is younger than 543.9 ± 0.2 Ma. On the other hand, data from Siberia (Brasier *et al.*, 1994b; Kaufman *et al.*, 1996; Kouchinsky *et al.*, 2001) show that the Nemakit-Daldynian Stage is lacking intervals with low negative $\delta^{13}C$ values characteristic of Member A-4 in the Ara Group and with uniform values of +3.0 ± 0.5‰ typical of Member A-5. To eliminate these contradictions, it is necessary to check the C-isotope system retentivity in carbonates of the Fara Formation and Ara Group and to carry out additional investigations.

It would be pertinent as well to use data of Sr-isotope chemostratigraphy to substantiate correlations based on $\delta^{13}C$ variations, but this is impossible so far. First, the Sr-isotope parameters of the above considered sections are known fragmentary (the number of published Sr-isotope analyses is less than that of $\delta^{13}C$ values by an order of magnitude and more). Second, the $^{87}Sr/^{86}Sr$ initial ratios calculated for limestones and dolostones of the Khufai, Shuram, and Buah formations of Oman are inappropriate for chemostratigraphic correlation. The reason is that all the samples studied do not satisfy strict geochemical criteria accepted in this research, which identify the Rb–Sr system retentivity in carbonates, and the samples have not been preliminarily leached to remove their secondary carbonate phases what is an obligatory procedure as we believe (Kuznetsov *et al.*, 2003; Semikhatov *et al.*, 2003). Third, the $^{87}Sr/^{86}Sr$ initial ratios in "best" samples from the pre-Nemakit-Daldynian interval of the Upper Vendian are of a narrow variation range. For instance, these ratios in the lower part of the Zaris Formation and in upper strata of the Urusis Formation in Namibia correspond to 0.70840 and 0.70845–0.70858 respectively (Kaufman *et al.*, 1993), whereas in the lower part of the Ust-Yudoma Formation in Siberia they range from 0.70829 to 0.70844 (Semikhatov *et al.*, 2003). Such a low gradient of secular changes excludes a possibility to use these ratios for a proper correlation in this part of geological record.

CONCLUSIONS

Inferences from data considered in this work are as follows.

(1) Representative C-isotope characteristics are first obtained for the Lower Yudomian carbonates from two sections of the Yukanda Formation in the Yudomian stratotype area. The least altered carbonate samples from the Yukanda Formation, which elucidate further and extend downward C-isotope chemostratigraphy of the Yudomian stratotype formerly studied in its upper interval, were selected based on strict geochemical criteria ($Mn/Sr \le 4$, $Fe/Sr \le 10.0$, $\delta^{18}O > -10‰$ in limestones and $Mn/Sr \le 6$, $Fe/Sr \le 15$, $\delta^{18}O > -10‰$ in dolostones).

(2) In samples selected under control of the above criteria, $\delta^{13}C$ values demonstrate the in-phase variations of comparable amplitude at the correlative levels of both studied sections. They are used to plot the generalized curve of $\delta^{13}C$ variations in the Lower Yudomian succession that is divided, based on respective trends, into intervals corresponding to three subformations of the Yukanda Formation. Characteristic of the Yalan Subformation, the lower one, are prevailing high positive $\delta^{13}C$ values ranging from +6 to +8‰ and outlining a brief excursion toward lower values near the base of the upper one fourth, whereas near the top they are within the range of +4.9 to +3.2‰. In the Mal Subformation, this parameter declines down to moderate positive values with two short negative excursions (+0.8 and –1.2‰) complicating the general trend. Low negative $\delta^{13}C$ values are typical of the Tokur Subformation (the range from –7.8‰ near the base to –6.8‰ near the top).

(3) C-isotope chemostratigraphy of Yudomian successions in the Olenek Uplift and stratotype area of the unit, which are reliably correlated at the Nemakit-Daldynian level based on biostratigraphic criteria, implies that the Yukanda (Aim) Formation of the Lower Yudomian and the Ust-Yudoma Formation of the Upper Yudomian are separated by a considerable hiatus in the Uchur–Maya region. The hiatus spanning the accumulation interval of the Maastakh, Khatyspyt, and, probably, lower Turkut deposits in the Olenek Uplift was not less than 4 to 4.5 m.y., as one can judge from correla-

tion with concurrent deposits of Namibia, where a series of U–Pb dates was obtained for volcanogenic zircons.

(4) The presented correlation of Yudomian deposits in the Olenek Uplift and Uchur–Maya region determines the Siberian pre-Nemakit-Daldynian succession of terminal Precambrian intervals with specific C-isotope signatures. This succession not only corresponds the standard one suggested by Saylor *et al.* (1998) but also extends the latter downward by the interval of Yalan Formation, which antedates the standard interval N_{pg}, being different from it in characteristic $\delta^{13}C$ values. The Yalan Subformation and its equivalents, e.g., the Khufai Formation of Oman with comparable stratigraphic position and C-isotope parameters, should be discriminated into the separate chemostratigraphic interval that can be termed the Varangerian positive (V_p). It is convenient to accept the levels of $\delta^{13}C$ inversion around zero, which are recorded in the lower Varangerian cap carbonates and at the base of interval N_{pg}, for the limits of the interval in question.

(5) C-isotope chemostratigraphic correlation between the Vendian reference sections in Siberia, Namibia, and Oman, which has been controlled by biostratigraphic and isotopic-geochronological data, implies that their pre-Nemakit-Daldynian successions are divisible into five separate intervals (V_p, N_{pg}, P_r, P_f, and I), in which $\delta^{13}C$ secular variations are of individual character and amplitude.

(6) Relatively high positive $\delta^{13}C$ values in basal horizons of the Yalan dolostones suggest that these horizons are stratigraphically higher than cap carbonates of the lower Varangerian level, characteristic of which are negative $\delta^{13}C$ values rising upward in the section (Kaufman *et al.*, 1997; Walter *et al.*, 2000; Brasier *et al.*, 2000; and references therein). Consequently, the carbonate part of Yudomian succession in the stratotype area is lesser in time range than the Vendian with its lower boundary corresponding to the base of the lower Varangerian tillites.

(7) It is impossible to judge on age relations between basal siliciclastic deposits of the Lower Yudomian (Nachar Subformation and its equivalents) and the lower Varangerian tillites, although some indirect data suggest that these deposits accumulated at the time of the early Varangerian deglaciation and corresponding sea-level rise.

(8) Short-term negative $\delta^{13}C$ excursions down to values of –1.2 to –2.5‰, which are recorded in the Mal Subformation of the Uchur–Maya region and near the Khufai Formation top in Oman, suggest, along with stratigraphic position of both subdivisions, their possible connection with events of the late Varangerian glaciation, although this conclusion needs in additional confirmations.

ACKNOWLEDGMENTS

We are grateful to M.A. Fedonkin and I.M. Gorokhov for a fruitful discussion of problems considered in the work. Assistance of A.B. Kotov and D.P. Vinogradov in determination of C- and O-isotope compositions in some samples, and analytical work of I.V. Kislova, who measured Mn and Sr concentrations in all the samples, are greatly appreciated. The work was supported by the Priority Program no. 6 of the Earth Science Division, Russian Academy of Sciences, and by the Russian Foundation for Basic Research, project nos. 02-05-64333 and 00-05-72011.

Reviewers I.M. Gorokhov and M.A. Fedonkin

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Stratigraphy and Geological Correlation, Vol. 12, No. 5, 2004, pp. 460–484. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 12, No. 5, 2004, pp. 29–54.
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Boundary between Permian Series in Diverse Sedimentary Facies of North European Russia: Constraints of Event Stratigraphy

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Received March 6, 2003; in final form, July 13, 2003

Abstract—Lithologic-facies and ecologic-taphonomic methods are jointly used to study the Lower–Upper Permian boundary deposits of a complex structure, which are represented by laterally variable marine, coastal, deltaic, lagoonal, and terrestrial facies. The created landscape model elucidates spatial relations between different sedimentary facies and respective sea-level fluctuations in the Kungurian shallow basin. The established brachiopod zonation is carefully correlated with ammonoid biostratigraphy. It is shown that mid-Kungurian and Solikamsk stratigraphic events are of a high correlation potential in the Boreal realm. The Kungurian age is substantiated for marine facies of the northeastern Pai-Khoi Ridge and Subpolar Urals, which are correlative with the Solikamsk Horizon. New results and data on interconnection between the lower Kazanian marine and Sheshma red-bed facies imply inexpediency of including the Ufimian Stage into the General stratigraphic scale. The Kazanian Stage base corresponding to the Lower–Middle Permian boundary in the International scale is most appropriate candidate for the General scale level that should separate the Lower Permian from the upper or middle series in the two- or three-unit subdivision of the system, respectively.

Key words: Permian System, Kungurian, Ufimian and Kazanian stages, brachiopod zonation, ammonoids, biofacies, sedimentation settings, sea-level fluctuations, event stratigraphy.

INTRODUCTION

Debates on boundaries and ranges of the Kungurian and Ufimian stages, and respectively on the Lower–Upper Permian boundary lasted several decades. The main problem was the range of the Ufimian Stage whose lower boundary had not been reliably substantiated. In original definition, the Ufimian Stage corresponded to the "lower red-bed sequence" between the Kungurian and Kazanian deposits. The sequence range changed several times, as it has been regarded at times as including red beds of the Kungurian, Kazanian, and even Tatarian stages in their current understanding (Forsh, 1955). The "lower red-bed sequence" got the stage status, when Nechaev (1915) simultaneously attributed the overlying gray marine deposits to the Kazanian Stage (Table 1). In the General stratigraphic scale, that sequence is authorized merely in the rank of the Ufimian Formation (*Resolution of the Permian...*, 1950), the lower boundary of which was defined based on lithological criteria without paleontological characterization. At that time, the Solikamsk Formation was included into the Kungurian Stage. In 1960, the Ufimian Formation was renamed (*Stratigraphic Schemes...*, 1962) and attributed to the Sheshma Horizon (*Resolutions of the Conference...*, 1962). The latter and Solikamsk Horizon were jointly included into the lower subdivision of the Upper Permian as the legalized Ufimian Stage (*Resolutions of the Conference...*, 1962). In the stratotype area, paleontological characterization of Solikamsk Horizon is controversial and poor. The lower boundary of this horizon (and correspondingly of the Ufimian Stage) is defined at the appearance level of lycopods *Viatcheslavia vorcutensis* Zal. and their phylloids *Viatcheslaviophyllum vorcutense* Neub. Other organic remains, predominantly freshwater fossils, are of a local significance, unsuitable for substantiation of the stage lower boundary outside the stratotype area. During the long-lasted debates on the Solikamsk Horizon range, Likharev (1968) attributed it to the Kungurian Stage and argued: "its fauna inhabited a basin that was tightly connected with the Early Permian Cis-Uralian sea or a relict sea basin as a result of a short-term transgression" (p. 177). It is also problematic how the Kungurian and Ufimian stages of the General scale should be coordinated with the Kungurian and Roadian stages of the International scale (Table 1).

Consequently, it is necessary to specify ranges of the Kungurian and Ufimian stages in marine sequences of the Boreal realm in order to define boundary between Permian series of the general scale. It is necessary therefore to rank the distinguishable stratigraphic events, to trace the reference correlation levels of the Kungurian–Kazanian interval, and to correlate the lat-

Table 1. Correlation scheme for subdivisions distinguished by different authors in the Lower–Upper Permian boundary interval

General stratigraphic scale (GSS)				Regional stratigraphic scheme of Russian platform	Nechaev, 1915		Resolutions of the Permian conference, 1950		Resolutions of Conference of 1962		Changes in the GSS (proposal)	International stratigraphic scale
System	Series	Stage		Horizon								
Permian	Upper	Kazanian	upper	Povolzhskii	Kazanian Stage		Kazanian Stage		Kazanian Stage	upper	Kazanian Stage	Wordian
			lower	Sok						lower		Roadian
		Ufimian		Sheshma	Ufimian Stage	Lower red beds	Ufimian Formation		Ufimian Stage	Sheshma Horizon		
				Solikamsk	Kungurian Stage		Kungurian Stage	Solikamsk Formation		Solikamsk Horizon	Kungurian Stage	Kungurian Stage
	Lower	Kungurian*		Iren				Iren Formation	Kungurian Stage	Iren Horizon		
				Filippovo				Filippovo Formation		Filippovo Horizon		
				Sarana	Artinskian Stage		Artinskian Stage		Artinskian Stage			

*Before 1998, the Sarana Horizon was in the Artinskian Stage (ISC resolutions of 1998).

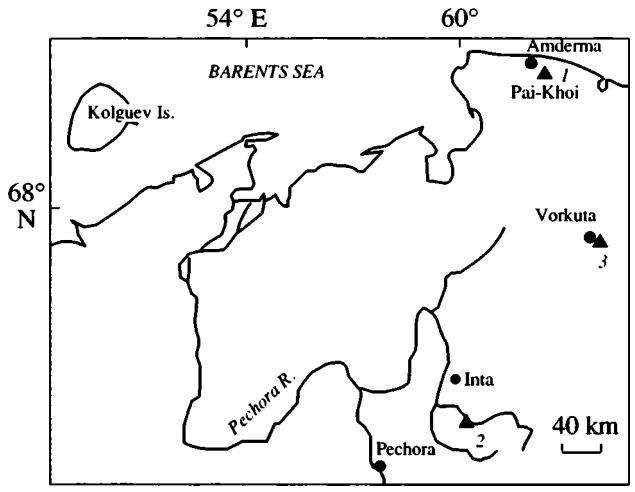


Fig. 1. Geographic localities of the Tab-Yu (*1*), Kozhim (*2*) and Vorkuta (*3*) sections studied in this work.

ter with boundary stages of the Lower–Middle Permian of the International scale.

MATERIALS AND METHODS

In 1999–2001, we described in detail the most complete marine and lagoonal-continental sequences of Kungurian–Ufimian deposits in the northeastern Pai-Khoi Ridge (Tab-Yu River), Polar (Vorkuta River) and Subpolar Urals (Kozhim River) (Fig. 1). The bed-by-bed description was carried out with parallel sampling of organic remains.

The materials obtained clarified and updated the paleontological characterization of studied sections. The established brachiopod zonation was meticulously correlated with distribution of macroflora and index taxa of ammonoids.

As the studied sequences are composed of laterally variable sedimentary facies, we had to apply a complex of lithologic-facies methods to analyze influence of landscape factors on biotic diversity of fossil organisms and to understand the nature of distinguishable biostratigraphic subdivisions. In addition, these methods elucidated secular trend of eustatic sea-level fluctuations, which are useful for remote correlation and for identification of superregional geological events.

Based on genetic typification of lithologic beds (Frolov, 1984), we analyzed the recurrent sedimentary cycles to reconstruct the lateral system of facies tracts characterizing shallow shelf and coastal zones of a basin with terrigenous sedimentation in a belt of temperate humid climate. The landscape-ecologic model created for that basin of the Polar Urals (Figs. 1, 2) was used to figure out the sea-level variation curve for each of the sections studied (Figs. 3, 4). Based on the curves, we analyzed frequency and intensity of transgressive and regressive phases in order to define records of high-rank landscape reorganizations, which could be related to the eustatic sea-level fluctuations.

Data on changes in brachiopod taxonomic diversity and taphonomic analysis of faunas and floras were used to define dominant and accessory components of fossil assemblages. The recurrent appearance of discriminated biofacies was an additional criterion for recognition of eustatic fluctuations.

The used methods of zonal biostratigraphy and facies analysis provided grounds for the substantiated correlation of the sections studied. As a result, we determined levels of the Kungurian–Ufimian interval, which are most significant in terms of event stratigraphy of the Polar and Subpolar Urals.

STRATIGRAPHY

Northeastern Pai-Khoi Ridge. The most complete section of marine and lagoonal–continental Kungurian–Ufimian deposits is exposed in the Tab-Yu middle canyon of the northeastern Pai-Khoi Ridge (Figs. 1, 3). In 1972–1973, that section was studied in detail by geologists from the Vorkuta Complex Geological Expedition and Komi Filial of the All-Union Research Institute of Gas (Entsova *et al.*, 1974), who divided the Lower–Upper Permian boundary deposits into the Talota, Tab-Yu, and Er-yaga formations. The first of them was renamed afterward into the Liur-yaga Formation (Pukhonto, 1998; Kuklev *et al.*, 2000). The Tab-Yu Formation lower boundary was defined in the section at the appearance level of coastal facies with characteristic brachiopod species *Lingula arctica* Miloradovich. The upper one was placed at the first occurrence level of plant remains typical of the "Pechora" flora (Entsova *et al.*, 1974). Later on however, deposits of Liur-yaga Formation were included into the Tab-Yu Formation as it follows from paleontological characteristics presented below (Gus'kov *et al.*, 1980; Kanev and Kalashnikov, 1990).

Liur-yaga Formation. The formation stratotype is in exposures L-14 and L-15 of the Liur-yaga upper canyon, NE Pai-Khoi Ridge (Pukhonto, 1998). It has been described in brief by Kanev and Kalashnikov (1990), Pukhonto (1998), and Kuklev *et al.* (2000). In the stratotype, the formation is composed of alternating gray sandstone and siltstone beds with rare shale interlayers, which are free of coal seams and identifiable plant remains. Geologists of OAO "Polyarnouralgeologiya" and Kalashnikov (Kanev and, Kalashnikov, 1990) reported that the formation rocks yield brachiopods *Arctitreta kempei* (Anders.), *Sowerbina borealis* (Haugh.), *Megousia kuliki* (Fred.), *Anidanthus aagardi* (Toula), *Thuleproductus arcticus* (Whitf.), *Striapustula koninckiana* (Fred.), *Waagenoconcha humboldti* Orb., *W. irginae* Stuck., *Yakovlevia vorcutana* (Sol.), *Rhynchopora nikitini* (Tschern.), *Spiriferella rajah* Salt., *Sp. saranae* Vern., and *Sp. draschei* (Toula). Bogoslovskaya (1997) described ammonoids *Tumaroceras*

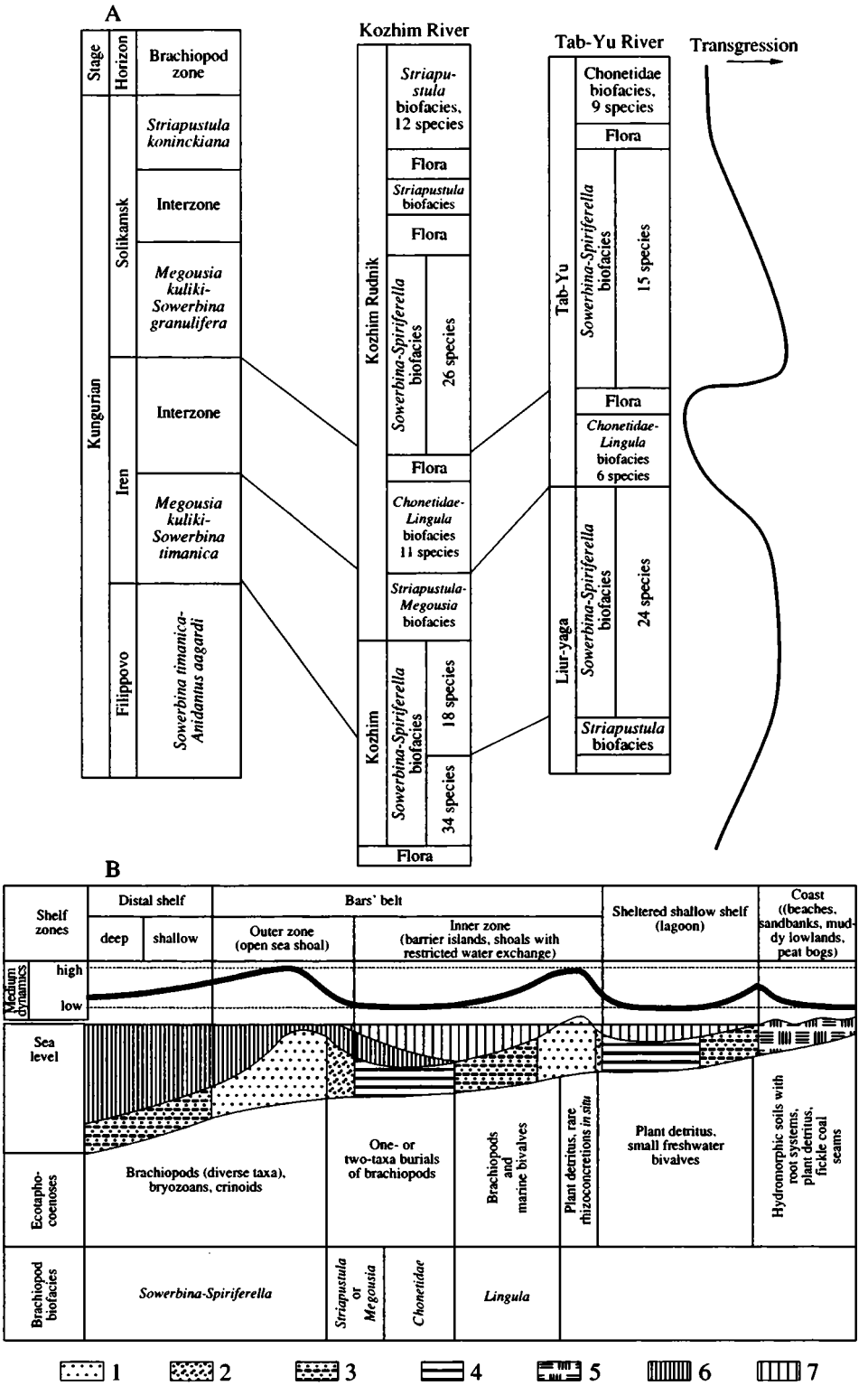


Fig. 2. Distribution of biofacies (A) in the Kozhim and Tab-Yu sections and landscape-ecological model (B) of coastal zones in a shallow basin of the northern Cis-Urals, the terminal Early Permian: (1) micro- to fine-grained sands; (2) clayey fine-grained sands with stirring and landslide marks; (3) alternating silt and fine-grained sand; (4) clay or silt; (5) sand, silt, clay and thin peat horizons; (6) normal and (7) lowered salinity.

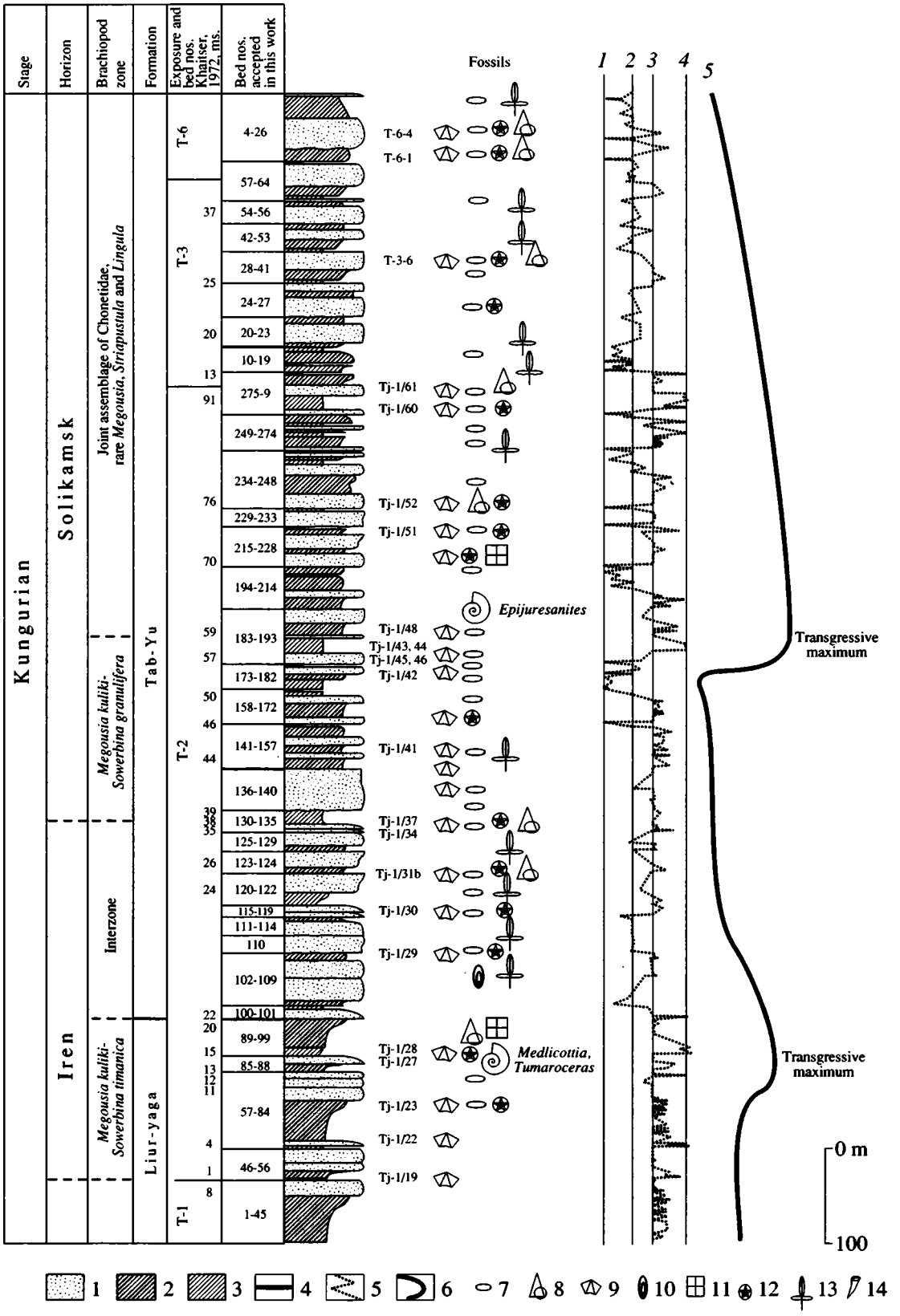


Fig. 3. Lithostratigraphy of Lower–Upper Permian boundary interval in the Tab-Yu section (northeastern Pai-Khoi Ridge): (1) sandstone; (2) siltstone; (3) shale; (4) coal seams; (5) sea-level fluctuation curve; (6) transgressive–regressive sedimentary cycles of high order; (7) bivalves; (8) gastropods; (9) articulate and (10) inarticulate brachiopods; (11) bryozoans; (12) crinoids; (13) plant remains; (14) solitary Rugosa corals; (*1*) coastline; (2) sheltered shallow shelf; (*3*) bars' belt; (*4*) distal shallow shelf; (*5*) distal deep shelf.

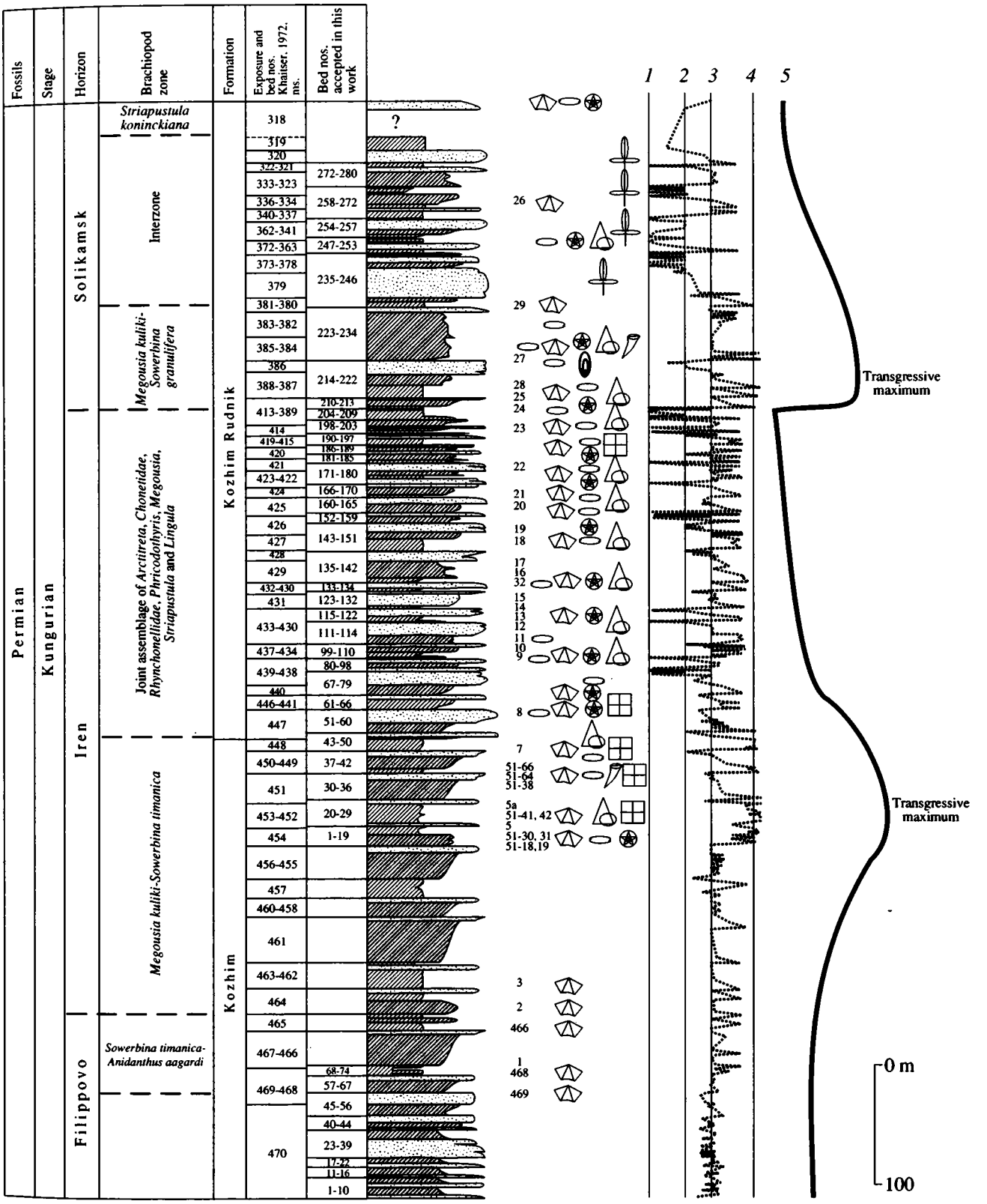


Fig. 4. Lithostratigraphy of Lower–Upper Permian boundary interval in the Kozhim section of the Subpolar Urals (symbols and other indications as in Fig. 3).

dignum Bogoslovskaya and *Medlicottia postorbiniana* Bogoslovskaya from the upper part of the Liur-yaga Formation stratotype, which substantiate the Kungurian age of the unit.

In the Tab-Yu middle canyon, the Liur-yaga Formation is represented by sandstone and siltstone beds 1–99 (Pukhonto, 1998; Kuklev *et al.*, 2000); the apparent thickness is 115 m (Fig. 4). The beds were originally attributed to the Talota Formation distinguished in exposures of the Bol'shaya Talota River, the southwestern flank of the Pai-Khoi Ridge (Gus'kov *et al.*, 1980).

The studied section consists of five lithologic cycles 10 to 40 m thick. Grain size of sedimentary material increases upward within each of the cycles. The cycle base is composed of dark gray silty shale (0/5 to 3 m) that is interpreted as sediment of a deep distal shelf. The greater middle interval (3 to 20 m) exhibits frequent lenticular–banded alternation of clay siltstone and fine-grained sandstone laminae. Frequency and thickness of sandstone laminae increase upward. This part of cycles likely corresponds to sediments deposited in shallow settings of an open shelf. In upper parts of the cycles, there are two- or three-unit members of fine-grained sandstone (up to 10 m) of the outer bar belt. The cyclic structure of the section suggests periodic fluctuations of the basin depth and recurrent transitions from relatively deep settings of an open shelf (basal intervals of cycles) to external shallow environments of a bars' field (upper parts of cycles). Hypertrophied lower and reduced upper parts of the last two cycles (beds 87–93) are indicative of the first transgressive maximum recorded in the section.

Abundant and diverse brachiopods are characteristic of the formation upper part (beds 46–99). The beds yield *Arctitreta kempei* (Anders.), *Svalbardia capitolina* (Toula), *Lissochonetes koejimensis* Mir., *Sowerbina timanica* (Stuck.), *Waagenoconcha irginae* (Stuck.), *W. wimani* (Fred.), *Kochiproductus saraneanus* (Fred.), *Anidanthus aagardi* (Toula), *Megousia kuliki* (Fred.) (mass specimens), *M. aurita* Sol. (mass specimens), *Thuleproductus arcticus* (Whitf.), *Anemonaria pseudohorrida* (Wim.), *Costatumulus* cf. *janischewscianus* (Step.), *Striapustula multicostata* Sol., *S. koninckiana* (Keys.), *Rhynchopora nikitini* Tschern., *Stenoscisma biplicata* (Stuck.), *Neospirifer permicus* Ifan., *Spiriferella draschei* (Toula), *Sp. vojnowski* Ifan., *Timaniella vasjagensis* (Ifan.), and *Phricodothyris asiatica* (Chao). The assemblage is Kungurian in age, because its dominant taxa are known from Kungurian deposits of the Timan–Pechora region, Cis-Urals, Spitsbergen (base of the Kapp Starostin Formation, *"Horridonia" timanica* Zone), and northeastern Russia (top intervals of the Tumara and Dzhigdali horizons, *Megousia kuliki* Zone). Based on position in the section and similar brachiopod assemblages, the considered part of the section is correlated with an upper interval of the Talota Formation stratotype at the Bol'shaya Talota River, southwestern flank of the Pai-Khoi Ridge.

Tab-Yu Formation. The formation lower part about 500 m thick (the complete thickness is 800–900 m) is exposed in the Tab-Yu middle canyon (Entsova *et al.*, 1974; Pukhonto, 1998). According to original definition, its lower boundary is at the level of transition from sediments of a distal sea shelf to the complex of coastal facies deposited in a regressing basin with a changeable, predominantly low salinity.

In the formation basal interval (beds 100–140, 130 m thick in total), there are 12 lithologic cycles. Two-unit bar cycles (2 to 15 m thick) with a granulometric minimum near the top (thicker basal sandstone and thinner upper part of alternating siltstone and fine-grained sandstone laminae) are most common. In the upper half cycles, there are three-unit deltaic cycles (10 to 25 m thick) with a granulometric minimum near the base (silty shale or clay siltstone of avandelta below, alternating siltstone and sandstone of delta front in the middle, and sandstone of the delta plain in the greater upper interval). In general, the formation lower part was deposited within the bars' field at the regressive stage of sedimentation.

Organic remains occurring at the base of the Tab-Yu Formation are impoverished, different in composition from underlying fossils. *Lingula arctica* Milor. and *L. liurjakhensis* Kalash. represent lingulids, which first appear here, become dominants higher, and coexist with rare chonetids and rhynchonellids *Rhynchopora nikitini* (Tschern.). Gastropods and marine bivalves *Aviculopecten orientalis* Fred., *Astartella permocarbonica* Tschern., *Pyramus? symmetricus* (Lutk. et Lob.), and *Permophorus?* sp. are widespread. Species *Paracalamites* cf. *striatus* Schm., *P.* sp. (*P.* aff. *frigidus* Neub.), *Sphenopteris* sp., *Pecopteris* ex gr. *borealis* Zal., *Cordaites* sp., and Anthracosidae forms are identified among plant remains from lower beds 102–109. In the middle interval (beds 118–120), *Paracalamites frigidus* Neub., *P.* ex gr. *frigidus* Neub., *Cordaites* ex gr. *singularis* (Neub.), and *C.* sp. indet. occur in association with single *Viatcheslavia* sp. (*V.* cf. *vorcutensis*) first appearing at this level. Plant assemblage of this kind has been interpreted as an atypical Solikamsk flora (Naugol'nykh, 1998). Beds 131–136 yield diverse brachiopods *Arctitreta kempei* (Andrus.), *Waagenoconcha wimani* (Fred.), *Striapustula multicostata* Sol. [*St.* cf. *koninckiana* (Keys.)], *Megousia kuliki* (Fred.), *M. aurita* (Sol.), *Sowerbina granulifera* (Toula), *Linoproductus rhiphaeus* Step., and *Yakovlevia vorcutensis* Sol. whose assemblage is close to that from the Liur-yaga Formation. Nevertheless, it marks the first occurrence and prevalence of *Sowerbina granulifera* (Toula) that replaced *Sowerbina timanica* (Stuck.). Both the Liur-yaga and Tab-Yu brachiopod assemblages include species characteristic of the Kungurian *Megousia kuliki* Zone in the Boreal realm.

The overlying 220 m of the section (beds 141 T-2–19 T-3) are divisible into 11 lithologic cycles of a complex structure. Characteristic of each cycle is the following complete or reduced succession of sedimentary facies replacing one another from the base upward:

1. Sandstone of the bars' belt outer zone, which yields remains of marine fauna;

2. Frequently alternating laminae of clayey and coarse-grained siltstones deposited in distal shelf settings and bearing sea fossils;

3. Sandstone of the bars' belt inner zone, in which plant remains are buried;

4. Frequently alternating laminae of clayey and coarse-grained siltstones deposited in sheltered shelf settings and bearing bivalves and floral remains;

5. Coastal siltstone with coal seams and root remains suggesting impact of pedogenic processes.

The cycles are 10 to 30 m thick. Their sequence was likely deposited under influence of recurrent transitions from distal shelf settings to bars, lagoons, and coastal sedimentation areas. Beds 183–193 characterize a cycle that is 30 m thick and composed almost entirely of the lower distal-shelf facies, while the upper ones are reduced. This cycle apparently recorded a maximum of the middle Tab-Yu transgression that commenced the new large transgressive–regressive cycle of sedimentation after the regression of the early Tab-Yu time.

Beds 145–186 yield brachiopods *Arctitreta kempei* (Anders.), *Dyoros grandis* Afan., *D. pseudotrapezoidalis* (Mil.), *Sowerbina granulifera* (Toula), *Waagenoconcha wimani* (Fred.), *Striapustula multicostata* Sol., *Megousia kuliki* (Fred.), *M. aurita* Sol., *Rhynchopora nikitini* (Tschern.), *Neospirifer permicus* Ifan., *Spiriferella draschei* (Toula), and *Sp. vojnowski* Ifan. In the Bed 186, *Kochiproductus* sp., *Waagenoconcha* sp., *Megousia kuliki* (Fred.) and *?. aurita* Sol. occur in association with *Epijuresanites* sp. Rather abundant bivalves are represented by *Aviculopecten orientalis* Fred, (small forms), *Astartella permocarbonica* Tschern., *Stutchburia* cf. *polassi* (Vern.), and *Praeundulomia* cf. *gijigensis* (Masl.). Higher, within the interval from Bed 187 (exposure T-2) to Bed 19 (exposure T-3), the brachiopod assemblage is less diverse, predominantly consisting of lingulids (*Lingula artiensis* Mil., *L. hyperborean* Ifan.), chonetids [*Arctochonetes postartiensis* (Ustr.), *Lissochonetes rotundatus* (Toula), *L. koejimensis* Mir., *Svalbardia capitolina* (Toula)], and rhynchonellids (*Rhynchopora nikitini* Tschern.). Rare *Striapustula* sp., *Strophalosia* sp., and *Phricodothyris* sp. occurring here are transit forms present up to the Tab-Yu Formation top. The concurrent bivalve assemblage is of a uniform composition, consisting of single *Pseudomonotis* cf. *speluncaria* (Schlot.) and *Polidevcia* cf. *flexuosa* (Lutk. et Lob.) and of abundant *Astartella permocarbonica* Tschern. Plant remains in the upper half sequence are in contrast more diverse. Beginning from Bed 200 and higher, up to the formation top, they are represented by *Viatcheslavia vorcutensis* Zal. and *Viatscheslaviophyllum vorcutense* Neub. characteristic of the Inta Formation. These taxa occur in association with *Sphenophyllum* cf. *thonii* Mahr., *Paracalamites frigidus* Neub., *Sciadisca petschorense* Zal., *Phyllotheca striata* Schm., *Tschernovia striata* Neub., *Rufloria* cf. *ensiformis* (Zal.) S. Meyen, *R.* cf. *recta* (Neub.) S. Meyen, *Cordaites lineatus* (Neub.) S. Meyen, *C. singularis* (Neub.) S. Meyen, *Pecopteris compta* Radcz., *P. borealis* Zal., *Samaropsis vorcutana* Tschirk., *S. pogorevitschi* Dombr., *S. postfrigida* Dombr., *Cardiocarpus? minimus* Dombr., and *Intia* sp.

The upper part of the section exposed in the Tab-Yu middle canyon (Bed 20 of exposure T-3 to Bed 26 of exposure T-6) is up to 130 m thick and consists of five lithologic cycles. Sandstone members of these cycles are thicker than below (up to 15 m), while siltstone layers and coal seams are less frequent that seems to be a result of a more intense influx of clastic material at the regressive phase of sedimentation. Broad sand beaches and shoals were likely formed in the coastal zone during that phase. Bivalves *Polidevcia* cf. *flexuosa* (Lutk. et Lob.), *Astartella* cf. *permocarbonica* Tschern., and *Streblopteria* cf. *bella* Gusk. represent the dominant group of fossils. Concurrent brachiopods taxa are *Megousia* sp., *Strophalosia* sp., abundant *Rhynchopora* sp., and chonetids. Plant assemblage is represented by bark remains of lepidophytes *Viatcheslavia vorcutensis* Zal. and *Viatscheslaviophyllum vorcutense* Neub. associated with *Zamiopteris* ex gr. *glossopteroides* Schmalh., *Pecopteris borealis* Zal., *Samaropsis postfrigida* Dombr., *Rufloria* sp., *Paracalamites frigidus* Neub., *Paracalamitina striatus* (Schmalh.) Zal., and Anthracosidae.

Entsova *et al.* (1974) and Kuklev *et al.* (2000) reported on occurrence of workable coal beds in the upper sequence of the Tab-Yu Formation outside the canyon exposures. The coals deposits likely reflect an intense peat accumulation along coasts of large freshwater lagoons, which have been formed at the terminal stage of the late Tab-Yu regression.

The Tab-Yu Formation is correlated with the Solikamsk Horizon based on similar assemblages of marine and freshwater bivalves (Gus'kov *et al.*, 1980; *Biota of the East...*, 1998), which include the following species in common: *Cyrtorostra laticostata* Netsch., *Promytilus sicarius* Lutk. et Lob., *Astartella raricostata* Demb., *A. permocarbonica* Tschern., *Aviculopecten orientalis* Fred., *Vorcutopecten talis* Lutk. et Lob. (marine taxa), and *Khosedaella alta* (Pog.) and *Kh. rotunda* (Pog.) (non-marine). The inferred Solikamsk age is consistent with the Inta-type plant remains present in the upper half of the formation.

The Subpolar Urals, Kozhim River. Permian deposits of the region have been described first by Rotai (1946) who divided their sequence into the Kos', Chernaya Rechka, Kozhim, and Kozhim Rudnik formations, the last two subdivisions attributed to the Kungurian

Stage. During the subsequent reexamination (Kuz'kokova *et al.*, 1980; *Permian System...*, 1991; *Guide Book...*, 1995; *Biota of the East...*, 1998), nomenclature and ages of subdivisions changed several times. The Ufimian age suggested for the Kozhim Rudnik Formation (Murav'ev *et al.*, 1986, *Guide Book...*, 1995) is authorized in the Stratigraphic Scheme of the Urals of 1994. The Ufimian Stage lower boundary defined at the base of Bed 47 of the Kozhim Rudnik Formation is proposed as candidate for the respective GSSP (*Biota of the East...*, 1998; Grunt *et al.*, 1999). In 1999–2001, we studied the Kozhim and Kozhim Rudnik formations in exposures of the Kozhim River (Fig. 4).

Kozhim Formation about 450 m thick rests conformably on the Chernaya Rechka Formation of clayey to silty deposits of a deep shelf, which are of the late Artinskian age. The lower part (about 100 m) of the Kozhim Formation (Bed 470 after Rotai) consists of asymmetric two-member sandy–aleuritic lithologic cycles ranging in thickness from 0.5 to 2.5 m. The greater lower member is composed of fine- to medium-grained sandstone with wavy lamination. The upper member with a gentle wavy lamination consists of alternating siltstone and fine-grained sandstone laminae. Characteristic of the rocks are fucoids oriented at random, coalified plant detritus, and ripple marks. Sediments are barren of macrofauna. Taxa identified among their plant remains are *Paracalamites* ex gr. *frigidus* Neub., *Cordaites* sp., *Rufloria* (*Alatorufloria*) cf. *recta* (Neub.) S. Meyen, *Bardocarpus aliger* Zal., and *Cordaicarpus anceps* Dombr. All the features suggest that deposits under consideration accumulated within a bars' belt during the terminal regressive phase of the Chernaya Rechka sedimentation cycle.

The overlying sequence about 250 m thick (beds 469–452 after Rotai) consists of lithologic cycles (10 to 25 m thick) similar in structure to the Liur-yaga cycles described above. In their lower thicker parts, the cycles are composed of lenticular alternating laminae of clay siltstone and highly calcareous fine-grained sandstone. Laminae of the last type become thicker and more frequent in the upward direction. Near their top, the cycles are composed of calcareous fine-grained sandstone 0.5 to 2 m thick. The section interval in question corresponds to deposits of an open shallow shelf zone and signifies commencement of the "Kozhim" transgression. The rocks yield abundant brachiopods *Derbyia grandis* Waag., *Dyoros pseudotrapezoidalis* (Mil.), *Thuleproductus arcticus* (Whitf.), *Reticulatia uralica* (Tchern.), *Krotovia tundrae* (Fred.), *Sowerbina timanica* (Stuck.), *Chaoiella volkovi* (Fred.), *Anemonaria pseudohorrida* (Wim.), *Anidanthus aagardi* (Toula), *Stenoscisma mutabilis* (Tschern.), *Paeckelmanella* cf. *dieneri* (Tschern.), *P.* cf. *expansa* (Tschern.), *Neospirifer* cf. *subfasciger* (Lich.), *Spiriferella saranae* (Vern.), *Sp. draschei* (Toula), and *Pinegathyris* ex gr. *amdrupi* (Dunb.). The assemblage is Kungurian in age as it follows from stratigraphic position of the Kozhim Formation conformably resting on the Chernaya Rechka Formation that bears the late Artinskian ammonoids *Paragastrioceras subtrapezoidale* Max. et Tch., *Uraloceras unicum* Bogosl. et Shkol. near the top and *Waagenina subinterrupta* (Krot.) 100 m below (*Biota of the East...*, 1998).

The Kozhim Formation upper part about 60 m thick (beds 453–448 after Rotai) consists of three deltaic lithologic cycles (15–20 m thick), clastic material of which is getting coarser upward through the cycle sequence. The Kozhim deltaic cycles have been already considered in detail (Shishlov and Zhuravlev, 2001). In distinction from overlying cycles, they show the well-developed lower interval of fine-grained sediments, which were deposited in prodelta settings at the initial stage of the delta front advancing. These widespread facies of distal shelf correspond to the first "Kozhim" transgressive maximum. The indicated upper beds of the Kozhim Formation bear brachiopods *Rhipidomella* cf. *permiana* Kul., *Dyoros pseudotrapezoidalis* (Mil.), *Neochonetes adzvensis* Ifan., *Thuleproductus arcticus* (Whitf.), *Th. subarcticus* Sar., *Sowerbina timanica* (Stuck.), *Reticulatia uralica* (Tschern.), *Chaoiella volkovi* (Fred.), *Anemonaria pseudohorrida* (Wim.), *A. bicarinata* (Wim.), *Linoproductus rhiphaeus* (Step.), *Anidanthus* sp., *Megousia kuliki* (Fred.), *M. aurita* Sol., *Yakovlevia impressa* (Toula), *Muirwoodia mammatiformis* (Fred.), *Striapustula koninckiana* (Keys.), *Costatumulus janischewskianus* (Step.), *Krotovia tundrae* Fred., *K. psedoaculeata* Krot., *Kochiproductus porrectus* (Kut.), *Waagenoconcha wimani* (Fred.), *W.* aff. *irginae* (Stuck.), *Bathymionia adzvensis* Kalash., *Rhynchopora variabilis* (Stuck.), *Pseudosyringothyris karpinskii* Fred., *Cyrtella kulikiana* Fred., *Neospirifer rhomboidalis* Kalash., *Kaninospirifer kaninensis* (Lich.), *Spiriferella draschei* (Toula), *Arcullina mica* (Barch.), and *Cleiothyridina semiovalis* (Fred.). The assemblage is successive with respect to older one, being more diverse in taxonomic aspect. In addition to endemics, it includes species widespread in the Kungurian Iren Horizon of the Cis-Urals and in the Vyl' Formation of North Timan (*Megousia kuliki* Zone of northeastern Asia).

Kozhim Rudnik Formation. Being 550–600 m thick, this subdivision is overlain by the coaliferous Inta Formation, although the direct contact between them is unexposed. In distinction from underlying unit, the Kozhim Rudnik Formation encloses coal seams (5 to 15 cm thick), hydromorphic paleosol horizons, and burials of plant remains. Its lower part about 50 m thick (beds 447–439 after Rotai) is represented by three deltaic cycles composed of delta-plain sediments (Shishlov and Zhuravlev, 2001), which characterize an initial stage of the "Kozhim Rudnik" regression.

The higher 250-m-thick sequence (beds 438–400 after Rotai) is composed of slightly coaliferous cycles, which are comparable in structure and genesis with middle cycles of the Tab-Yu Formation of the northeastern Pai-Khoi Ridge. This part of the Kozhim Rud-

nik Formation was deposited under conditions of open shelf, bar, lagoonal, and coastal settings recurrently changing during the terminal early–initial late "Kozhim Rudnik" regressive–transgressive cycle. Untypical brachiopod species of genera *Lingula, Arctitreta, Chonetes, Rhynchopora, Megousia, Striapustula*, and *Phricodothyris*, which occur in the lower interval of the Kozhim Rudnik Formation (beds 447–400 after Rotai), are of a low diversity. Beds 437–434 yield plant remains *Psygmophyllum expansum* Brong., *Rufloria (Alatorufloria)* ex gr. *recta* (Neub.) S. Meyen. A more representative assemblage of *Paracalamites frigidus* Neub., *P.* ex gr. *frigidus* Neub., *P. striatus* (Schm.) Zal., *Cordaites singularis* (Neub.) S. Meyen, *Rufloria* sp., *Zamiopteris lanceolata* (Chachl. et Pol.) Neub., *Z.* ex gr. *lanceolata* (Chachl. et Pol.) Neub., *Cardioneura vorcutensis* Zal., *Samaropsis frigida* Neub., and *Cardiocarpus* sp. was found in Bed 413.

The middle coal-free member is 80 m thick (beds 389–381 after Rotai), composed of shale and siltstone beds, which were deposited predominantly in deep settings of a distal shelf. These beds corresponding to the maximum of the late "Kozhim Rudnik" transgression yield a diverse brachiopod assemblage. Species of the assemblage are *Arctitreta triangularis* (Wim.), *A. macrocardinalis* (Toula), *A. kempei* Anders., *Dyoros pseudotrapezoidalis* (Mil.), *Lissochonetes spitzbergianus* (Toula), *Svalbardia capitolina* (Toula), *Craspedalosia pulchella* Dunb., *Waagenoconcha* cf. *irginae* (Stuck.), *Kochiproductus porrectus* (Kut.), *Megousia kuliki* (Fred.), *Thuleproductus arcticus* (Whitf.), *Yakovlevia impressa* (Toula), *Muirwoodia mammatiformis* (Fred.), *Sowerbina granulifera* (Toula), *Chaoiella volkovi* (Fred.), *Anemonaria bicarinata* (Wim.), *Striapustula koninckiana* (Keys.), *Rhynchopora nikitini* Tschern., *Cyrtella kulikiana* (Fred.), *Paeckelmanella dieneri* Tschern., *Spiriferella draschei* (Toula), *Sp. vojnowski* Ifan., *Eliva talbeica* (Ifan.), *Arculina mica* (Barch.), *Spirelytha borealis* (Mir.), and *Phricodothyris asiatica* (Chao). In distinction from the late Kozhim assemblage, among the listed species there are newcomers *Svalbardia capitolina, Craspedalosia pulchella*, and *Sowerbina granulifera* whose appearance does not influence, however, the age determination. Among smaller foraminifers occurring near the member top, there is *Gerkeina komiensis* Grozd. et Lebed., the index species of Kungurian deposits in Spitsbergen (Sosipatrova, 1981; Pronina, 1999).

The apparent thickness of the formation upper part (beds 380–318 after Rotai) is about 200 m. Lithologic cycles at this level are similar to those described above and include again some coal seams. In these cycles with well-developed pedogenic horizons, the distal shelf facies are extremely rare. The cycles accumulated in periodically changing settings (bars, lagoons, coastal areas) and characterize development of the late "Kozhim Rudnik" regression.

Characteristic of these deposits are dwarfed chonetids, rhynchonellids, *Striapustula* and *Megousia* forms, abundant small *Phricodothyris* species, and single *Waagenoconcha* specimens. Brachiopods are associated with bivalves, rare gastropods, and crinoids occurring in lenticular accumulations. According to Pukhonto (1998) and Naugol'nykh (1998), the Kungurian (Iren) flora of deposits includes the following taxa: *Paracalamites frigidus* Neub., *P. decoratus* (Eichw.) Zal., *Paracalamitina* cf. *striata* (Schmalh.) Zal., *Pecopteris* (Asterotheca?) kogimensis Zal., *P.* cf. *borealis* Zal., *Zamiopteris* sp., *Rufloria (Alatorufloria)* cf. *recta* (Neub.) S. Meyen, *Lepeophyllum* cf. *sabanakovae* Vlad., *Cordaites?* sp., *Xiphophyllum?* sp., *Samaropsis?* aff. *complanata* Naug., *Cardiocarpus* aff. *cordatus* (Eichw.) Schmalh., and *Sylvella?* sp. At the apparent top of the formation, there is a sandstone bed deposited likely in an outer bar zone (Bed 318 after Rotai). The bed yields the upper brachiopod assemblage of *Arctitreta kempei* (Anders.), *Svalbardia capitolina* (Toula), *Waagenoconcha wimani* (Fred.), *W. irginae* (Stuck.), *W.* sp., *Bathymionia adzvensis* Kalash., *Striapustula koninckiana* (Keys.), *S.* cf. *elongata* Gan. et Laz., *Megousia kuliki* (Fred.), *Kaninospirifer kaninensis* (Lich.), and *Neospirifer permicus* Ifan. This assemblage dominated by *Striapustula koninckiana* is less diverse than the older one from beds 210–234.

The greater part of the Kozhim Rudnik Formation (up to Bed 362 after Rotai in exposures of the Kozhim River) is correlated with the Solikamsk Horizon (*Biota of the East...*, 1998). It yields smaller foraminifers characteristic of the Kungurian *Gerkeina komiensis* and *Frondicularia prima* zones (Pronina, 1999). Among concurrent brachiopods, the most significant taxa are *Striapustula koninkiana* (mass abundance), *Megousia kuliki*, and *Sowerbina granulifera*. In opinion of Pukhonto (1998) and Naugol'nykh (1998), fossil flora also suggests the Kungurian age for the greater interval of the formation.

Polar Urals, the Vorkuta River. The Kungurian–Ufimian deposits are attributed here to the Lek-Vorkuta Formation that is 800 to 1500 m thick. Its lower part (Ayach-yaga Subformation 200 to 800 m thick) is comparable with the Tab-Yu Formation of the northeastern Pai-Khoi Ridge and with the Kozhim Rudnik Formation of the Subpolar Urals, as one can judge from lithologic-facies and taphonomic indications and from the character of sedimentary cycles with coal seams. Mass accumulations of *Striapustula koninckiana* (Keys.) and associated chonetids, *Neospirifer permicus* Ifan., *Arctitreta kempei* (Anders.), *Linoproductus* cf. *ripheaus* (Step.), and *Waagenoconcha* sp. are buried in rare sandstone interlayers of subformation, which were deposited in the outer bar zone.

The upper Rudnitskaya Subformation 150 to 600 m thick encloses main workable beds of the Vorkuta coal deposit, being composed of classical bar–lagoonal lithologic cycles (Makedonov, 1965). The subforma-

tion similar to the overlying Inta Formation was deposited at the final stage of the "Lek-Vorkuta" regression. Member "M" that separates two subdivisions is 100 m thick and corresponds to the transgression maximum of the terminal Lek-Vorkuta time. It is composed of less coaliferous deposits of the sea-coastal settings. Coquinas of *Waagenoconcha* brachiopod shells are buried in sandstone interlayers deposited in an outer bar zone.

Brachiopods of the Lek-Vorkuta Formation indicate the Kungurian age of host deposits that is consistent with opinion of Vorkuta geologists (Pukhonto, 1998; Kuklev *et al.*, 2000).

BRACHIOPOD ZONATION

In the Tab-Yu River section (Figs. 2, 3; Tables 2, 3; Plate I), we distinguished the *Megousia kuliki–Sowerbina timanica* Zone (beds 46–99), an interzone (beds 100–130), and *Megousia kuliki–Sowerbina granulifera* Zone (beds 131–186), which are of a complex substantiation. The selected index species *Sowerbina timanica* and *S. granulifera* are genetically interrelated taxa successively replacing one another in the section.

The lower boundary of *Megousia kuliki–Sowerbina timanica* Zone corresponds to the first occurrence level of index species associated with *Striapustula koninckiana* (Keys.), abundant *Megousia aurita* Sol., and *Striapustula? multicostata* Sol. The lower boundary of *Megousia kuliki–Sowerbina granulifera* Zone marks the appearance and subsequent abundance of *Sowerbina granulifera* and *Dyoros* species. According to general taxonomic composition and dominant species, this zone is close to the *Megousia kuliki–Sowerbina timanica* Zone. Ammonoids *Epijuresanites* sp. found at the top of *Megousia kuliki–Sowerbina granulifera* Zone (Kotlyar *et al.*, 1999) determine its Kungurian age. Ammonoids are confined to Bed 186 at the level of 350 m above the formation base and 200 m higher than the first occurrence level of *Viatcheslavia* forms.

Brachiopods of the *Megousia kuliki–Sowerbina timanica* and *Megousia kuliki–Sowerbina granulifera* zones are very diverse, of the mixed Ural–Pechora–Kolyma type. In interzones, they are much less diverse, represented by rare dwarfed *Megousia* forms, chonetids, rhynchonellids, and lingulids. These impoverished assemblages are of persistent composition up to the Tab-Yu Formation top.

In the Kozhim River section, we established six biostratigraphic subdivisions (Figs. 2, 4; Tables 2, 4): *Sowerbina timanica–Anidanthus aagardi* (beds 469–465 after Rotai) and *Megousia kuliki–Sowerbina timanica* (beds 464–448 after Rotai) zones; an interzone (beds 447–440 after Rotai); *Megousia kuliki–Sowerbina granulifera* Zone (beds 389–381 after Rotai); one more interzone (beds 380–319 after Rotai); and *Striapustula koninckiana* Beds (Bed 318; see Tables 2, 3). Discrimination criteria and substantiation of boundaries were as those used in the Tab-Yu River section of the northeastern Pai-Khoi Ridge.

Brachiopod assemblages of *Sowerbina timanica–Anidanthus aagardi, Megousia kuliki–Sowerbina timanica,* and *Megousia kuliki–Sowerbina granulifera* zones are diverse, characterizing transgressive phases of sedimentary cycles. In interzones, where brachiopods are of a low diversity, dominant fossils are represented by bivalves whose shells are buried in coquina interlayers and lenses. These taphonomic changes indicate the basin shoaling at the regressive phases of cycles. The impoverished brachiopod assemblages consist of chonetids, rhynchonellids, thin-shelled lingulids, and smaller *Striapustula* forms.

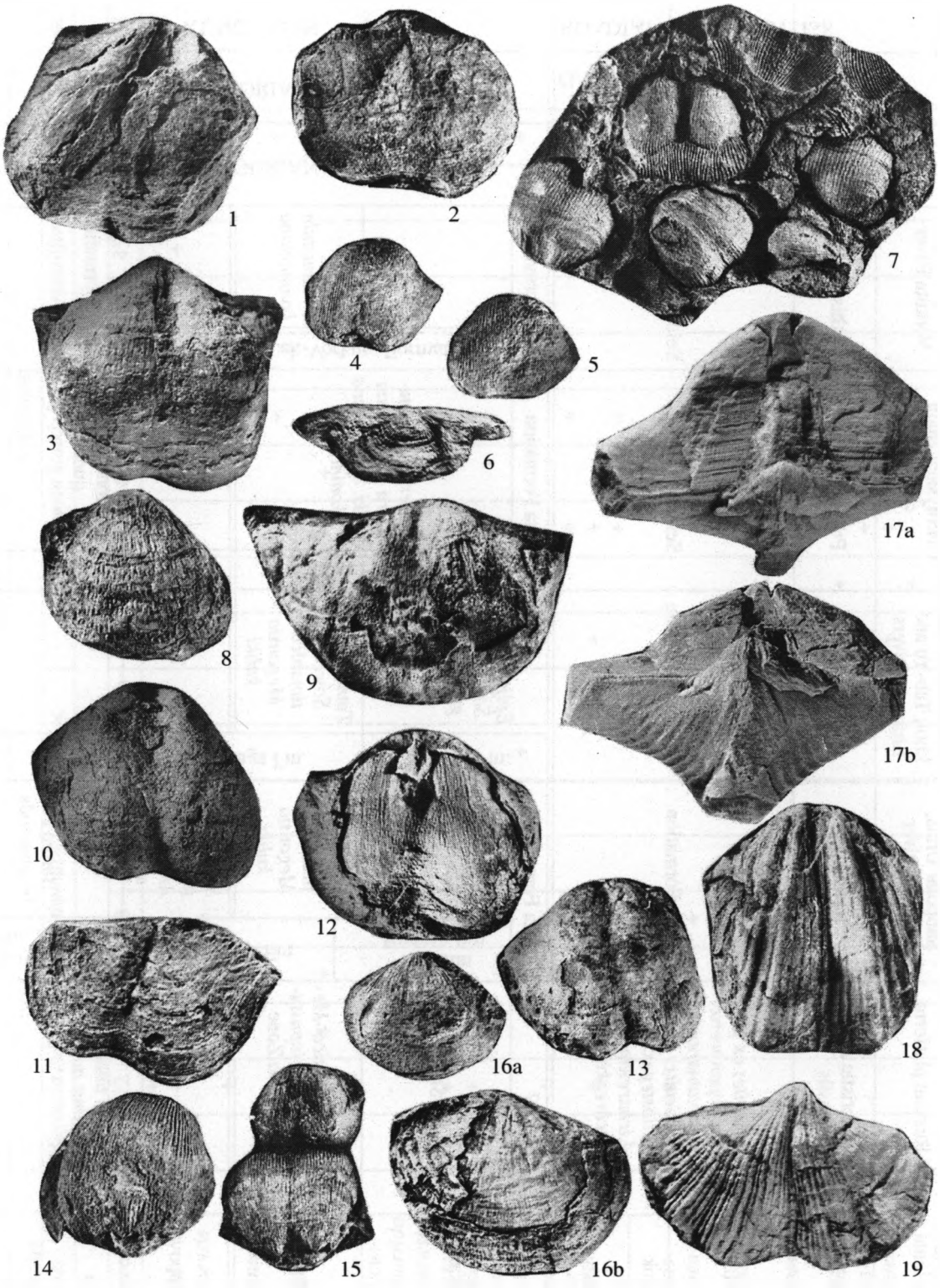
The *Megousia kuliki–Sowerbina timanica* and *Megousia kuliki–Sowerbina granulifera* zones of the Kozhim and Pai-Khoi regions are well correlative. The *Megousia kuliki–Sowerbina timanica* Zone is well traceable throughout the Boreal province up to far northeastern Asia, where it was distinguished first as the *Megousia kuliki* Zone in the upper part of the Dzhigdali Superhorizon of Kolyma–Omolon massif (Ganelin, 1973; Ganelin *et al.*, 2001) and in the Tumara Horizon of the Verkhoyansk region (Klets *et al.*, 1998).

The *Megousia kuliki–Sowerbina granulifera* Zone (middle part of the Tab-Yu and Kozhim Rudnik formations) bearing at the top *Epijuresanites* forms is of the Kungurian age. Consequently, the base of Bed 447 in the Kozhim River section cannot be regarded as boundary between series of the Permian System as it was suggested before (*Biota of the East...*, 1998).

Characteristic brachiopod species of *Sowerbina timanica–Megousia kuliki* and *Sowerbina granulifera–Megousia kuliki* zones.

Sowerbina timanica–Megousia kuliki Zone: (1–3) *Sowerbina timanica* (Stuckenberg, 1875), Tab-Yu River, Liur-yaga Formation, Bed 58 (1, 2) and Kozhim River, Kozhim Formation, Bed 2 (3); (4–6) *Megousia kuliki* (Fredericks, 1915), Kozhim River, Kozhim Formation, Bed 2 (4, 5) and Tab-Yu River, Liur-yaga Formation, Bed 60 (6); (7) *Megousia aurita* (Solomina, 1957), Kozhim River, Kozhim Formation, Bed 468; (8) *"Striapustula" multicostata* (Solomina, 1957), ×1.5, Tab-Yu River, Liur-yaga Formation, Bed 46.

Sowerbina granulifera–Megousia kuliki Zone: (9) *Yakovlevia vorcutensis* (Solomina, 1957), Tab-Yu River, Tab-Yu Formation, Bed 130; (10–13) *Sowerbina granulifera* (Toula, 1875), Kozhim River, Kozhim Formation, Bed 213 (10) and Tab-Yu River, Tab-Yu Formation, Bed 41 (11–13); (14, 15) *Megousia kuliki* (Fredericks, 1915), Kozhim River, Kozhim Formation, Bed 213 (14) and Tab-Yu River, Tab-Yu Formation, Bed 150 (15); (16) *"Striapustula" multicostata* (Solomina, 1957), Tab-Yu River, Tab-Yu Formation, Bed 41, without exaggeration (16) and ×1.5 (16b); (17a, 17b) *Cyrtella kulikiana* (Fredericks, 1916), Kozhim River, Kozhim Formation, Bed 213; (18) *Spiriferella draschei* (Toula, 1875), Tab-Yu River, Tab-Yu Formation, Bed 150; (19) *Neospirifer permicus* Ifanova, 1972, Kozhim River, Kozhim Formation, Bed 213.


1
2
7
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17a
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17b
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16a
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16b
19

Table 2. Correlation scheme of sections in the North European Russia

GSS			Regional scale of Russian platform		Subpolar Urals, Kozhim River		Northeastern Paikhoi, Tab-Yu and Liur-yaga rivers		Chernyshev chain of hills		Vorkuta River		Renewed GSS	TSS	ISS
System	Series	Stage	Horizon	Characteristic zonal fossils	this work		this work		Pukhonto, 1998		this work				
Permian	Upper	Kazanian	Povolzhskii	*Sverdrupites* ex gr. *harkeri*, *Popanoceras* ex gr. *subtumarense*, *Neouddenites* (?), *Aulosteges*, *Licharewia schrenkii*	Seida Formation		Er-yaga Formation		Seida Formation		Seida Formation		KAZANIAN	MURGABIAN	WORDIAN
			Sok											KUBERGANDIAN	ROADIAN
		Ufimian	Sheshma												
			Solikamsk	?	Inta Formation		Tab-Yu Fm.	*Epijuresanites Sowerbina – granulifera-Megousia kuliki*	Inta Formation		Inta Formation		KUNGURIAN	BOLORIAN	KUNGURIAN
				Bocharella–Striapustula	Kozhim Rudnik Fm.	*Striapustula*			Adz'va Formation	*Sowerbina granulifera – Megousia kuliki, Striapustula koninckiana*	Lek-Vorkuta Formation				
						Sowerbina granulifera-Megousia kuliki									
	Lower	Kungurian	Iren	*Sowerbina timanica–Megousia kuliki–Striapustula koninkiana* Zone	Kozhim Fm.	*Sowerbina timanica-Megousia kuliki*	Liur-yaga Fm.	*Tumaroceras-Sowerbina timanica – Megousia kuliki*				*Striapustula koninckiana*			
			Filippovo	*Sowerbina timanica – Anidanthus aagardi*		*Sowerbina timanica – Anidanthus aagardi*		?		Siltstone–shale sequence	Talota Formation				
			Sarana		Chernaya Rechka Fm. (top)	Barren of brachiopods									

Note: Abbreviations GSS, TSS and ISS denote General, Tethyan and International stratigraphic scales.

Table 3. Stratigraphic ranges of brachiopods in the Tab-Yu River section

Stage	Kungurian																		
Horizon	Iren						Solikamsk												
Zone or Beds	*Megousia kuliki–Sowerbina timanica*						Interzone			*Megousia kuliki–Sowerbina granulifera*									Interzone
Formation	Liur-yaga						Tab-Yu												
Brachiopod species	19	22	23	27	28	28a	29	31c	31b	34	37	41	41a	42	43	44	45	46	48	51-T-6-4
Artitreta kempei		+								+		+		+						
Svalbardia capitolina		+					+			+						+	+			+
Lissochonetes koejimensis				+	+	+	+												+	
Sowerbina timanica		+		+																
Waagenoconcha irginae				+																
W. wimani				+						+		+		+					+	
Kochiproductus saraneanus					+														+	
Megousia kuliki		+	+	+						+		+	+	+					+	
M. auritus		+	+							+		+	+						+	
Thuleproductus arcticus				+																
Anemonaria pseudohorrida				+																
Costatumulus janischewskianus				+																
Striapustula multicostata	+	+								+		+								
Striapustula koninckiana		+								+				+						
Yakovlevia vorcutensis												+								
Rhynchopora nikitini							+	+			+								+	+
Stenoscisma biplicata		+																		
Neospirifer permicus												+		+						
Spiriferella. draschei		+		+	+					+		+							+	
Spiriferella vojnovski				+	+							+		+						
Spiriferella saranae																				
Timaniella vasjagensis		+																		
Phricodothyris asiatica					+												+			+
Lingula arctica							+	+												+
Lisochonetes liurjakhensis							+													
Chonetidae							+	+	+											
Dyoros pseudotrapezoidalis										+		+								
Dyoros grandis												+								
Sowerbina granulifera										+	+	+								
Striapustula elongata										+		+								
Linoproductus rhiphaeus										+		+								
Lingula hyperborea											+									+
Lissochonetes rotundatus																+	+			+
Arctochonetes postartiensis																+	+			+
Strophalisia sp.																				+
Striapustula sp.																				+

Table 4. Stratigraphic ranges of brachiopods in the Kozhim River sections

Stage	Kungurian													
Horizon	Filippovo					Iren								
Zone	*Sowerbina timanica–Anidanthus aagardi*					*Sowerbina timanica–Megousia kuliki*								
Formation	Kozhim													
Brachiopod species \ Beds	469	468	467	466	465	464	51–18	51–19	51–31	51–33	51–42	51–43	51–64	51–66
Derbyia grandis	+	+		+		+								
Derbyia simensis					+									
Dyoros pseudotrapezoidalis	+	+				+					+			
Thuleproductus arcticus	+	+			+	+								
Reticulatia uralica	+													
Krotovia tundrae	+	+		+	+	+	+	+	+	+	+			
Yakovlevia duplex						+								
Sowerbina timanica	+	+	+	+		+			+	+		+		
Chaioella volkovi	+	+	+			+		+			+		+	
Anemonaria pseudohorrida	+	+		+	+	+			+		+			
Anidantus aagardi	+	+		+										
Waagenoconcha aff. *irginae*		+		+						+			+	
Stenoscisma spitzbergiana	+				+									
Stenoscisma mutabilis	+		+											
Rhynchopora nikitini	+	+		+	+	+			+		+		+	
Paeckelmanella dieneri	+	+	+											
P. expansa		ex gr.												
Neospirifer subfasciger		cf.		+	+									
Neospirifer rhomboidalis						+								+

Table 4. (Contd.)

Brachiopod species \ Beds	469	468	467	466	465	464	51–18	51–19	51–31	51–33	51–42	51–43	51–64	51–66
Spiriferella saranae	+		+											
Sp. draschei	+					+		+	+	+		+		+
Pinegathyris amdrupi		+												
Rhipidomella permiana						+								
Kochiproductus porrectus						+			+				+	
Arctitreta macrocardinalis						+								
Megousia kuliki						+			+	+	+		+	+
Anidanthus knyrkoi						+								
Linoproductus rhiphaeus						+								
Costatamulus janischewskianus						+								
Yakovlevia mammatiformis						+						+		
Thuleproductus subarcticus						+			+	+				+
Krotovia pseudoaculeata						+			+					
Anemonaria bicarinata						+								
Cyrtella kulikiana						+			+		+		+	+
Neospirifer rhomboidalis						+			+					
Arcullina mica						+					+			
Cleiothyridina semiovalis						+			+					
Thuleproductus crassauritus							+	+						
Pseudosyringothyris karpinskii							+	+						
Yakovlevia artiensis												+		
Kaninospirifer kaninensis													+	
Yakovlevia impressa													+	+
Neochonetes adzvensis													+	+

SEDIMENTATION SETTINGS AND BIOFACIES

During the mid-Permian time, biota of the North Uralian basin developed in a shallow zone under conditions of a temperate climate. Brachiopods usually represent the dominant group of fossils. Crinoids, bryozoans, marine and freshwater bivalves are less abundant and represent accessory components of biota. Stenohaline spiriferids are characteristic of the distal shelf deposits only, while productids occur in a wider spectrum of facies that may reflect their ability to adapt the changeable salinity of seawater. The analyzed diversity dynamics of brachiopods, evaluated dominants of other fossil groups, and taphonomic considerations have been used to depict vertical and lateral changes of biofacies (Fig. 2) and to establish the corresponding eustatic fluctuations.

In the Kozhim River section, there is recorded a transition from the bar-belt sedimentation settings (lower part of the Kozhim Formation) to the distal shelf facies (upper part of that formation) and then, via deltaic facies, to coastal settings isolated periodically by bars (Kozhim Rudnik Formation) and to facies of large freshened lagoons (Inta Formation). In the Tab-Yu River section, the distal shelf facies (Liur-yaga Formation) grade upward into sediments of the bars' zone (lower part of the Tab-Yu Formation). Higher in the section, there is recorded alternation of distal and proximal shelf facies, which grade into lagoonal facies near the Tab-Yu Formation top (Kuklev *et al.*, 2000). The Ayach-yaga Subformation of Lek-Vorkuta Formation is interpreted as a facies analogue of the middle Tab-Yu and upper Kozhim Rudnik strata. The Rudnitskaya Subformation of Lek-Vorkuta Formation is composed of facies deposited in large freshened lagoons with marshy coasts (Makedonov, 1965). These lithologic peculiarities determine differences between concurrent biofacies in the indicated sections.

Uniformly distributed nest- and lens-shaped accumulations of remains of crinoid–pelecypod–brachiopod and bryozoan–crinoid–brachiopod (sometime with gastropods) communities are buried in the distal shelf facies (upper part of the Liur-yaga Formation) deposited under conditions of high to moderate hydrodynamic activity. The brachiopod *Sowerbina–Spiriferella* biofacies present here is of a maximum diversity. In this biofacies, productids and spiriferids occur in proportion somewhat variable along the strike, and the other groups are represented by rare specimens. It was difficult therefore to define dominant species of the biofacies. The same *Sowerbina–Spiriferella* biofacies of the Kozhim Formation lower part (*Sowerbina timanica–Anidanthus aagardi* and *Megousia kuliki–Sowerbina timanica* zones) originated in close landscape settings. In the course of transgression development, the species number increased from 18 to 34. Coexistence of bryozoans and crinoids whose remains are well preserved suggests a normal salinity of seawater. The taxonomic diversity maximum of biofacies is confined to the transgression peak.

Ecotaphocoenoses in the inner part of bars' belt consist of one or two taxa of smaller brachiopods associated with accessory crinoids and bivalves. The *Striapustula* or *Striapustula–Megousia* biofacies distinguishable here mark either the diversification at the transgression initial phase (lower part of the Liur-yaga Formation) or the diversity decrease at the commencement of regression (the Kozhim Rudnik Formation base). In addition, similar biofacies are characteristic of transgressive pulses recognizable within the general regressive trend (the Kozhim Rudnik Formation top).

The Chonetidae–*Lingula* biofacies is typical of freshened settings in the bars' inner belt. Taphocoenoses of this biofacies include lingulid, chonetid, and bivalve remains occurring in stratal or lenticular accumulations. The biofacies is characteristic of regressive cycles of the Tab-Yu, Kozhim Rudnik, and Lek-Vorkuta formations. The associated plant remains are those of lycopods (*Viatcheslavia* forms), cordaitanthaleans, and ferns.

Biofacies of brackish-water bivalves is characteristic of sediments deposited in closed shelf settings with freshened waters (final stages of regressive cycles of the Lek-Vorkuta, Kozhim Rudnik, and Tab-Yu formations). In this case, fossil flora is an individual component frequently dominating in taphocoenoses. Plant remains of coastal lagoonal zones represent calamitaleans, cordaitanthaleans, fern fronds, and well-preserved gymnosperm seeds. Newcomers are bryopsids. Coastal deposits of small lagoons with coal seams and hydromorphic paleosol horizons characterize the regression peaks recorded in the Kozhim Rudnik and Tab-Yu formations and in the Ayach-yaga Subformation of Lek-Vorkuta Formation. During short-term transgressive pulses, these ecotaphocoenoses were replaced by *Striapustula* remains, which were likely transported from the basin inner areas. The permanent lagoons were favorable for intense peat accumulation that gave rise to formation of workable coal beds in the Rudnitskaya Subformation of Lek-Vorkuta Formation, at the top of the Tab-Yu Formation, and in the Inta Formation.

Based on comprehensive analysis of biofacies we distinguished two well-traceable levels with diverse brachiopod assemblages, which are separated by intervals impoverished in brachiopods (Fig. 2A). The first *Sowerbina–Spriferella* biofacies corresponds to the base of *Megousia kuliki–Sowerbina timanica* Zone in the Kozhim Formation upper part and is traceable near the Liur-yaga Formation top in the Tab-Yu River exposures. The second *Sowerbina–Spiriferella* biofacies corresponds to the base of *Megousia kuliki–Sowerbina granulifera* Zone of the middle Kozhim Rudnik and Tab-Yu strata.

The *Striapustula* biofacies of brachiopods and specific burials of *Waagenoconcha* shells appear in the Vorkuta River section in addition to plant taphoc-

oenoses and Chonetidae–*Lingula* and brackish-water bivalve biofacies widespread in the lower part of Ayach-yaga Subformation and in Member "M" of Rudnitskaya Subformation of Lek-Vorkuta Formation. Their appearance was connected with two transgression peaks corresponding to two levels of *Sowerbina–Spiriferella* biofacies of the Tab-Yu and Kozhim rivers.

The two biofacies levels identified in remote sections based on lithologic-facies analysis imply the regional scale of transgressive maximums, which were likely related to eustatic sea-level fluctuations.

EVENT STRATIGRAPHY

The analyzed diversity dynamics of principal fauna groups and identified regional transgressive maximums show that in the Kungurian–Kazanian interval of studied sections there are recorded two stratigraphic events of a high correlation potential.

Mid-Kungurian event. A considerable biotic event took place in the mid-Kungurian time (Iren Horizon). The level of event has been studied in detail in exposures of the Tab-Yu (northeastern Pai-Khoi Ridge), Kozhim (Subpolar Urals), and Vorkuta (Polar Urals) rivers and has been traced as well in other regions (Tables 2, 5). Characteristic of the level was the taxonomic diversification of brachiopods (*kuliki–timanica* time), wide radiation of their Ural–Pechora–Kolyma communities, and development of taxonomic similarity over the entire Boreal realm. *Megousia kuliki* (Fred.), *Striapustula koninckiana* (Keys.), and *Sowerbina timanica* Stuck. became dominant in brachiopod assemblages. Concurrent bivalves demonstrate a wide distribution of the genus *Aphanaia* (*andrianovi* time), appearance of first Gondwanan immigrants, and taxonomic diversity maximum in the Permian period (Byakov, 2000). Genera *Ragozina, Praeundulomia, Vakunella, Pyramis, Merismopteria*, and others are occurring at the level for the first time (Muromtseva and Gus'kov, 1984). In community of smaller foraminifers, the mid-Kungurian event marks appearance of *Pseudonodasaria, Rectoglandulina*, and *Langella* forms (Pronina, 1999), species diversification of genera *Rectoglandulina* and *Frondicularia*, persistence of taxonomic composition over the entire Boreal realm, and universal presence of zonal index species *Gerkeina komiensis* Grozdilova et Lebedeva and *Frondicularia prima* Gerke. In comparison with the late Artinskian assemblage, diversity of ammonoids is somewhat lower, and their community was of a mixed type, as the Verkhoyansk–Kolyma genera *Baraioceras, Tumaroceras*, and *Epijuresanites* appeared in addition to Uralian *Paragastrioceras, Uraloceras, Talassoceras*, and *Metalegoceras*. Reorganization of the initial Iren time was related to an intense transgression.

Late Kungurian (Solikamsk) event recorded in the studied sections (Figs. 4, 5) gave rise to a higher taxonomic diversity of brachiopods of the *Megousia kuliki–Sowerbina granulifera* Zone, which are genetically interrelated with those of the *Megousia kuliki–Sowerbina timanica* Zone and characterize biogeographic radiation of last representatives of the Kungurian genus *Epijuresanites*. Species of this genus are known from analogues of the Solikamsk Horizon in the Vaigach Island (Bogoslovskaya, 1997), northeastern Pai-Khoi Ridge (Kotlyar *et al.*, 1999), Verkhoyansk area, Kolyma–Omolon massif, Kolyma River basin (Andrianov, 1985; Kutygin, 1999), and in southern Primor'e (Zakharov *et al.*, 1997). Foraminiferal assemblages with most characteristic *Saccamina arctica* Gerke and *Gerkeina komiensis* Grozd. et Leb. are of persistent composition throughout the Boreal realm. Pronina (1999) attributed these fossils to the uniform Iren-Solikamsk assemblage.

Our analysis showed that the event under consideration was connected with a short transgressive phase of eustatic character, as we think, which is clearly recorded in North European areas (Kozhim, Vorkuta, Tab-Yu Rivers, and Kolguev Island), Taimyr, Kolyma–Omolon massif, and Verkhoyansk region (Shishlov, 2003). That phase depicted in the sea-level variation curves influenced sedimentation environments in different landscape situations (Fig. 5). The event of Solikamsk transgression is an additional criterion for global correlation of sections in different biogeographic provinces. Earlier it was recognized in works by Ustritskii (1960) and Tikhvinskii (1977).

DISCUSSION

Based on the results obtained, it is possible to clarify the problem of stage boundaries in the General stratigraphic scale of Boreal realm.

The Kungurian Stage lower boundary defined at the appearance level of conodonts *Neostreptognathodus pnevi* from *Neostreptognathodus pequopensis–N. pnevi* phylogenetic lineage and corresponding to the base of Sarana Horizon (*Resolutions of the Interdepartmental...*, 1998) is traceable almost everywhere and coincides with the Kungurian boundary proposed for the International stratigraphic scale (Chuvashov *et al.*, 2002). It is sometimes problematic however to establish this boundary in Boreal realm despite the associated biotic events. In the Verkhoyansk and Kolyma–Omolon region, it is at the level, where Kungurian *Tumaroceras, Paratumaroceras*, and *Neouddenites* forms replace the Artinskian ammonoid assemblage. Conodonts *N. pnevi* present in the *Mesogondolella intermedia–Neostreptognathodus exculptus–N. pnevi* Zone of North America define position of the Kungurian lower boundary inside the Skinner Ranch Formation (Wardlaw and Glenister, 1995). In Tethyan areas, the boundary is at the base of *Mesogondolella intermedia–?. shindyensis* Zone correlated by Kozur (1995) with the *Mesogondolella intermedia–N. exculptus–N. pnevi* Zone of North America.

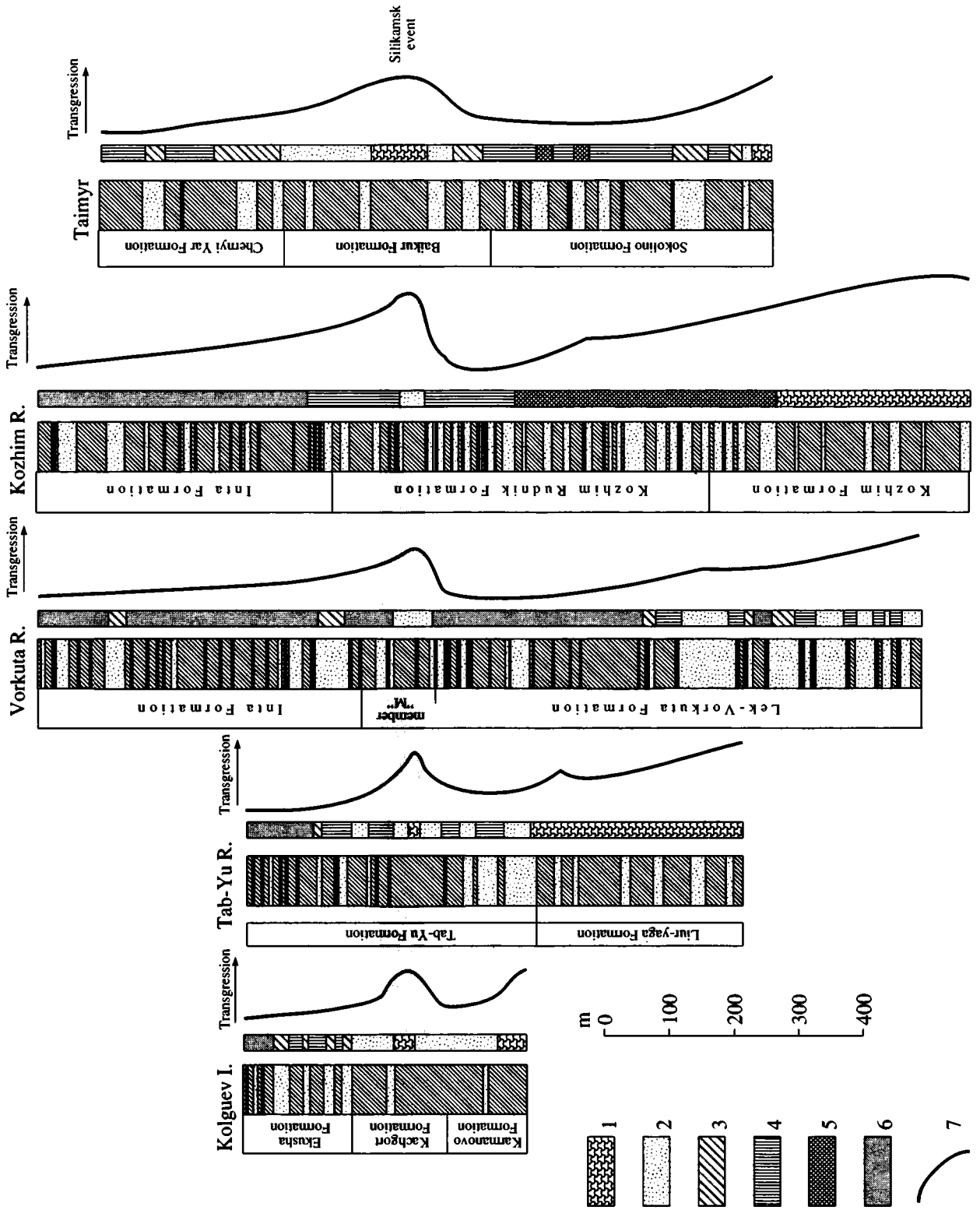

Fig. 5. North European (Kozhim R., Vorkuta, NE Pai-Khoi, Kolguev I.) and Taimyr sections correlated based on sea-level fluctuation curves and sedimentation settings: (1) deep shelf; (2) shallow shelf; (3) bars' belt; (4) sheltered shallow shelf; (5) deltaic tract; (6) coastal zone; (7) generalized curve of sea-level fluctuations (other symbols as in Fig. 3).

Table 5. Correlation scheme of sections in the North European Russia, Siberia, Far East, and Mongolia

GSS			RSS EEP	Kolguev Is.	Novaya Zemlya		Verkhoyansk region		Kolyma–Omolon massif		Southern Primor'e		Southern Mongolia		Renewed GSS (this work)	TSS	ISS
System	Series	Stage	Horizon	Entsova *et al.*, 1981; Konovalova, 1997	Povysheva and Ustritskii, 1996; Bogoslovskaya *et al.*, 1982		Klets *et al.*, 2001; Kutygin, 1999		Ganelin *et al.*, 2001		Kotlyar, 1997		Pavlova *et al.*, 1991; Manankov, 1998				
Permian	Upper	Kazanian	Povolzhskii	Ekushaan Fm.	Gerkino Fm.	*Sverdrupites amudseni*	Delendzha Horizon	*Sverdrupites amudseni*	Omolon Horizon	*Sverdrupites amudseni*	Vladivostok Horizon	*Daubichites*	Tsagaan-Uul Horizon	*Alispiriferella lita – Kaninospirifer adressum*	KAZANIAN	KUBERGANDIAN MURGABIAN	WORDIAN
			Sok		Kocherga Fm.	*Daubichites Sverdrupites harkeri*		*Daubichites Sverdrupites harkeri Mongolosia*		*Daubichites Sverdrupites harkeri Mongolosia*				*Liosotella decimana – Waagenoconcha angustata*			ROADIAN
		Ufimian	Sheshma														
			Solikamsk	Kachgort Fm.	Listvenichnaya Fm. Litkino Fm. Krotovo Fm.		Tumara Horizon	*Kolymaella ogonerensis*	Dzhigdali Horizon	*Kolymaella-Bocharella*	Abrek Horizon	*Primorewia reshetnikovi, Tomiopsis atlanichae*	Hovsgol Horizon	*Anidantus aagardi – Paramarginifera nativa – Neouddenites*	KUNGURIAN	BOLORIAN	KUNGURIAN
	Lower	Kungurian	Iren	Karmanovo Fm.				*Megousia kuliki Epijuresanites Tumaroceras*		*Megousia kuliki-Neouddenites*							
			Filippovo	?	Moller Fm.			*Anidantus aagardi*		*Anidantus aagardi*							
			Sarana														

Note: Abbreviations GSS, TSS and ISS denote the General, Tethyan and International stratigraphic scales, and RSS EEP is the Regional stratigraphic scale of the East European platform.

The mid-Kungurian event described above marks a higher level of the Iren Horizon base. The upper boundary of Kungurian Stage is defined in the General stratigraphic scale at the base of the Ufimian Stage represented by lagoonal continental facies in the stratotype and other areas of the Russian platform. Floras and freshwater faunas characteristic of these facies are unsuitable for discrimination of the Ufimian Stage and identification of its boundaries in marine sequences outside the type area. That is why similar assemblages of marine fossils have been regarded at times as a reasonable motivation to unite the Kungurian and Ufimian stages into a joint subdivision, e.g., into the Svalbardian (Stepanov, 1957) or Kamian Stage (Gorskii, 1974). Considering difficulties related to the Ufimian Stage discrimination, Ustritskii (1960) proposed to replace it by the Pai-Khoian Stage, but the new stage range had not been determined and it was unclear whether the suggested stratotype includes the Kungurian Stage as well.

The Solikamsk event corresponds to the *Megousia kuliki–Sowerbina granulifera* Zone in the upper half of Kozhim Rudnik Formation of the Subpolar Urals and to synonymous zone of the Tab-Yu Formation in the northeastern Pai-Khoi Ridge. Ammonoids present near the zone top define its Kungurian age. As it is generally accepted that this stratigraphic range is correlative with the Solikamsk Horizon of the Ufimian Stage, the horizon should be placed back into the Kungurian Stage, where its position was authorized long ago (*Resolutions of the Permian...*, 1950). Many experts on the Permian System argued for this position precisely in the middle of the last century (*Stratigraphic Dictionary...*, 1956; *Stratigraphy of the USSR...*, 1966; Stepanov, 1957; Tikhvinskaya, 1946). In view of the original range of the Kungurian Stage with the Solikamsk Horizon included, it is reasonable to retain the primary name of that stage.

If the Solikamsk Horizon is of Kungurian age, as substantiated above, then the Ufimian Stage corresponds in range to red beds of Sheshma Horizon only. The red beds (Ufimian Stage in terminology of A.P. Nechaev) are almost barren of fossils and "stratigraphic range of the Ufimian Formation is variably understood in different distribution areas of this unit" (Forsh, 1955, p. 26). Whereas some researchers united red beds of the Kungurian, Kazanian, and Tatarian (basal strata) stages (Krasnopol'skii, 1889; Stuckenberg, 1898), the others either considered them as facies equivalents of the lower *Spirifer* Limestone (Golovkinskii, 1869; Selivanovskii, 1935), or attributed partially to the lower Kazanian interval and partially to the Kungurian Stage (Tikhvinskaya, 1941). A doubtful paleontological characterization and difficulties in subdivision of the Sheshma red beds exclude a possibility to define reliably independence of the Ufimian Stage in nomenclature of Nechaev, its lower boundary, and affiliation with either the Lower or Upper Permian Series. Some indirect data suggest the Late Permian (Kazanian) level for a greater interval of the Sheshma red beds. N.A. Valeeva (see in Esaulova, 1996) and Igonin (1965) identified foraminifers *Pseudoammodiscus megasphaericus* (Gerke), *P. microsphaericus* (S. M.-Macl.), *Lingulonodosaria fallax* (Tcherd.), *Lingulina* cf. *semivelata* Tcherd., and *Ichtiolaria longissima* (K. M-Macl.) in rare marine interlayers of the Sheshma Horizon stratotype. In opinion of Pronina (1999), these species are characteristic of the *Nodosaria hexagona–Ichtyolaria subtilis* Beds in the lower part of the Kazanian Stage, and she attributed them together with foraminifers of the Baitugan Beds to a joint assemblage of the early Kazanian age. Esaulova (1998) who distinguished the Sheshma–Urzhum phylladoderm superassemblage also noted presence of Kazanian taxa in the Sheshma flora. Based on oldest therapsids of the Ocher assemblage (*Parabradysaurus silantjevi* tetrapod zone; Golubev, 1999), which have been found in cuprous sandstones of the Sheshma Horizon in the Perm region, the horizon was attributed to the Kazanian Stage. The Inta fauna (*Clamorosaurus nosturnus* tetrapod zone; Golubev, 1999) that antedates therapsids and is very similar to faunas of North America and Europe is attributed to the Lower Permian (Lozovsky, 2002). The presumable facies transition from lower Kazanian strata to Sheshma red beds found recently a concrete confirmation (Golubev, 2002; oral communication of Burov and Esaulova). At the same time, Forsh (1955, p. 26) reported that lower red beds, i.e., "first and second members of Ufimian strata exposed along the Belaya River correspond in age to the Kungurian Stage." If both viewpoints are admissible, then the greater upper part of red beds should be attributed to the Kazanian Stage and the lower one to the Kungurian stage. In addition, there is no place for the Sheshma Horizon in continuous marine sequences of the West Verkhoyansk and Kolyma–Omolon regions, if we regard it as a high-rank independent subdivision. The Ufimian Stage corresponds there to the *Kolymaella ogonerensis* Zone of Tumara and Dzhigdali horizons, which are correlated with the Solikamsk Horizon of the Russian platform, and deposits of that zone underlie Kazanian sediments with *Sverdrupites* and *Daubichites* (Klets *et al.*, 2001; Ganelin *et al.*, 2001).

Thus, it is clear that analogues of the Solikamsk Horizon are overlain by Kazanian sediments in southeastern marine sequences, and possible age equivalents of the Sheshma red beds cannot be discriminated here. In the sections we studied, there are continental deposits, which may be correlative with beds in question. As Tikhvinskaya and Likharev wrote, the Ufimian deposits could be regarded "as an individual stage, if there would be an intermediate sequence between the Kazanian and Kungurian stages in continuous sequence of marine strata, which would be distinct in paleontological characteristics from both stages. Situation like this is unknown so far" (*Stratigraphy of the USSR...*, 1966, p. 31).

Since recent data suggest the Kungurian age for the Solikamsk Horizon and the early Kazanian age for the greater interval of the Sheshma Horizon, it is impossible to regard the Ufimian Stage as a valid subdivision of the general stratigraphic scale.

Characteristic of the Kazanian Age commencement are remarkable changes in composition and structure of biota, first of all, the appearance of Roadian or Kazanian ammonoid assemblages instead of Kungurian predecessors (Tables 2, 5). The event (*Sverdrupites harkeri* time) corresponds to first occurrence level of genera *Sverdrupites, Pseudosverdrupites, Daubichites*, and *Anuites*. Simultaneously, majority of Kungurian taxa, except for rare *Popanocers, Neouddenites*, and *Medlicottia* became extinct. The most typical species of new assemblage are *Sverdrupites harkeri* (Ruzh.), *Pseudosverdrupites budnikovi* Kut, *Daubichites butakovensis* Bog., *D. gooshi* (Teich.), and *Anuites kosynskyi* Andr. The Roadian (Kazanian) assemblage that appeared almost instantly is traceable in all paleobiogeographic regions. In opinion of Chernykh *et al.* (2001), conodont species *Kamagnathus khalimbadzhae* Chernych from the base of the Baitugan Beds are very similar to *Sweetina triticum* Wardl. et Collin. from lower strata of the Road Canyon Formation of North America. Representatives of the genus *Kolymia* are dominants in the bivalve assemblage of the initial Kazanian time (Byakov, 2000). Dynamics of brachiopod diversification at this time is known inadequately, although it is clear that dominants of some phylogenetic lineages changed on another. For instance, *Megousia weyprechti* (Toula) replaced *M. kuliki*, genus *Terrakea* replaced genera *Bocharella* and *Kolymaella*, while genus *Sowerbina* and many other Kungurian taxa died out. New taxa, which became dominant in brachiopod communities are linoproductids with fine striation (*Terrakea* and *Spitsbergenia* forms) and genera *Mongolosia* and *Omolonia* widespread in the east of the Boreal realm (Ganelin *et al.*, 2001). The commencement of the Kazanian Age is marked by a wide expansion of diaphragmatophore corals. Near the base of the Kazanian Stage stratotype (Baitugan Beds at the Sok River), there is found "cyathoxonian fauna" characteristic of transgressive sedimentation cycles. The crinoid–bryozoan reefal buildups characteristic of temperate climate (quarries in the Nemda River basin) seem to be synchronous to this event. The *Calophyllum permianum* (Nechaev) assemblage from the Kamishlin Beds was likely typical of lagoonal facies. Corals of this assemblage radiated widely and existed later in basins of Greenland, Canada, and Zechstein.

Two unique ammonoid assemblages have been found in stratotype area of the Kazanian Stage: the lower one of *Sverdrupites* sp., *Popanoceras* ex gr. *subtumarense*, and *Neouddenites(?)* and the upper assemblage of *Sverdrupites* ex gr. *harkeri, Popanoceras* ex gr. *subtumarense, Medlicottia* sp., and *Altudoceras* sp.? or *Daubichites* sp. (Esaulova, 2001; Leonova *et al.*, 2002). They connect appearance of Roadian ammonoid assemblage with the commencement of the Kazanian Age. The ammonoids considered initially as typical of the upper Kazanian Substage only, occur, according to recent data, in the Baitugan and Kamyshlin beds of the lower substage as well (oral communication of Esaulova and Burov). Ammonoid assemblages studied by Leonova and Shilovskii are of a high correlation potential. It is also interesting that they include some Kungurian genera, and the successive character of Kungurian and Kazanian (Roadian) assemblages leaves therefore no place for an "intermediate" Ufimian epoch in evolution of ammonoids. Similar Roadian assemblages of ammonoids are known from the Novaya Zemlya Island, Kolyma–Omolon and Verkhoyansk regions, and in Primor'e, where the Kazanian Stage lower boundary is established respectively in the middle of Kocherga Formation and at the base of the Omolon, Delendzha, and Vladivostok horizons. Being defined at the appearance level of ammonoid species *Sverdrupites harkeri*, the lower boundary of the Kazanian Stage is confidently correlative with the base of the Roadian Stage of the International scale and of the Kubergandian Stage of the Tethyan scale (Tables 2, 5).

Global events characterize simultaneously the Kazanian Stage base and the Lower–Middle Permian boundary of the International scale. In the current General stratigraphic scale, this level can be regarded as the lower limit of the Upper Permian or, by transition to three-member scale, as the Lower–Middle Permian boundary.

CONCLUSION

(1) The landscape-ecological modeling of spatial relationships between different facies is used to reconstruct sea-level fluctuations of the Kungurian Age. Two regional transgressive events in a shallow basin of the northern Cis-Urals are distinguished.

(2) The established brachiopod zonation is comprehensively correlated with distribution of ammonoids and results of cyclostratigraphic analysis.

(3) The succession of biofacies is controlled, as is shown, by evolution of sedimentation environments. The *Sowerbina–Spiriferella* brachiopod biofacies of a high taxonomic diversity is an additional criterion for recognizing the transgression maximums.

(4) The mid-Kungurian and Kazanian stratigraphic events of a high correlation potential for the Boreal realm are recognized based on diversity dynamics of orthostratigraphic fossil groups and transgressive maximums established within the Kungurian–Kazanian interval.

(5) The analyzed biostratigraphic, sedimentological, and cyclostratigraphic data showed that the Solikamsk event, which is recorded in sections of North European Russia (Kozhim, Vorkuta, Pai-Khoi, Kolguev), Taimyr, Kolyma–Omolon massif, and the Verkhoyansk region,

was connected with a short-term transgression of eustatic origin.

(6) Equivalents of the Solikamsk Horizon among marine facies of the northeastern Pai-Khoi Ridge and Kozhim section of the Subpolar Urals are proved to be of the Kungurian age. As is established, the greater interval of the Sheshma red beds is correlative with the lower Kazanian marine strata, and it is unreasonable therefore to use the Ufimian Stage in the General stratigraphic scale.

(7) Our revision of available data implies that the Kungurian Stage includes the Sarana, Filippovo, Iren, and Solikamsk horizons and corresponds in range to its interval in the International stratigraphic scale.

(8) The Kazanian Stage base corresponding to the Lower–Middle Permian boundary in the International scale is recommended to be the lower limit of the Upper Permian in the current General stratigraphic scale, and by transition to three-member scale this level can be regarded as the Lower–Middle Permian boundary.

ACKNOWLEDGMENTS

The work was supported by the Russian Foundation for Basic Research, project nos. 99-05-65140, 01-05-64470, 03-05-65118, and by PalSPIR-Sepkoski Grants (2002).

Reviewers A.Yu. Rozanov and A.S. Alekseev

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Stratigraphy and Geological Correlation, Vol. 12, No. 5, 2004, pp. 485–494. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 12, No. 5, 2004, pp. 55–64.
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Maastrichtian–Paleocene Floras of Northeastern Russia and North Alaska and Floral Changes at the Cretaceous–Paleogene Boundary

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Received October 27, 2003

Abstract—Taphofloras of the Koryak Upland and the Sagwon Locality of North Alaska are described in terms of their taxonomic composition, stratigraphic ranges, and comparison with coeval floras of Northeastern Russia and Alaska. Causes and amplitude of changes in Arctic floras across the Cretaceous–Paleogene boundary are considered. In its aspect and taxonomic composition, the Late Maastrichtian flora of the Amaam Lagoon (the Koryak Upland) is very close to the early Paleocene Sagwon flora of North Alaska. This is inconsistent with a hypothesis of the global ecological crisis at the Cretaceous–Paleogene boundary. Development of floras was controlled by long-term climatic fluctuations and by evolution and migration of plants.

Key words: Cretaceous–Paleogene boundary, paleobotany, Northeastern Russia, North America.

INTRODUCTION

The Cretaceous–Paleogene boundary rightfully attracts attention of geologists and paleontologists, as it marks very dramatic events in history of the Earth's biota. Subjects of vivid discussions and debates are synchronism, stages, and magnitude of extinctions in various biotic groups in different regions of the globe and causes of this ecological crisis. For several decades the crisis was thought to have been caused predominantly by impact events (Alvarez *et al.*, 1980). The impact theory does not elucidate however a number of problems concerning the biota transit across this boundary. Meteoritic impacts were likely one of the factors that distorted ecological balance. It should be noted that principal evidences of catastrophic flora extinction in the terminal Cretaceous are known from mid-latitude North America (Spicer, 1989; Spicer *et al.*, 1994).

In this work, changes in Arctic floras at the Late Cretaceous–early Paleogene boundary time are considered based on floras known from below and above the boundary, i.e., on the late Maastrichtian flora of the Koryak Upland, Northeastern Russia, and on the early Paleocene Sagwon flora of North Alaska (Fig. 1).

THE KORYAK FLORA OF NORTHEASTERN RUSSIA

Continental and marine Upper Cretaceous–Lower Paleogene deposits are widespread in the Koryak Upland and further northward in the Anadyr' River basin and the Pekul'nei Ridge. Sequences in the Amaam Lagoon area, which are composed of alternating marine beds with mollusks and continental flora-bearing deposits, are of particular stratigraphic significance. The Koryak taphoflora is known from the upper Koryak Formation in the Amaam Lagoon area and the Emima–Il'naivaam interfluve (Fig. 2). Plant fossils first collected here by Volobueva and Terekhova (1974) were studied by Vasilevskaya and Abramova (1974) and later on by Golovneva (1994). The collection of fossils is stored under no. 1199 at the Botanical Institute of Russian Academy of Sciences (St. Petersburg). Plant remains collected by Golovneva and myself in the

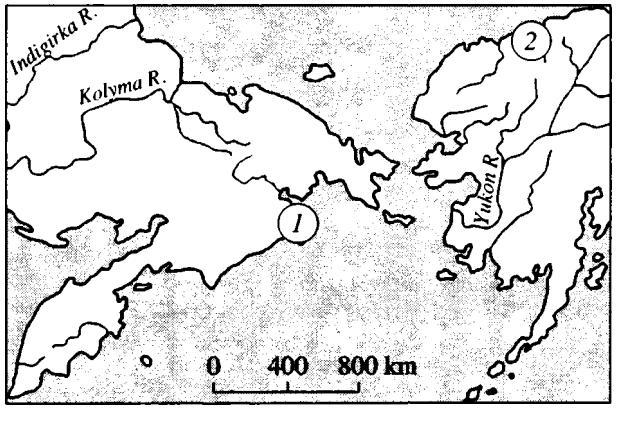

Fig. 1. Localities of the late Maastrichtian Koryak (*1*) and the early Paleocene Sagwon floras (*2*).

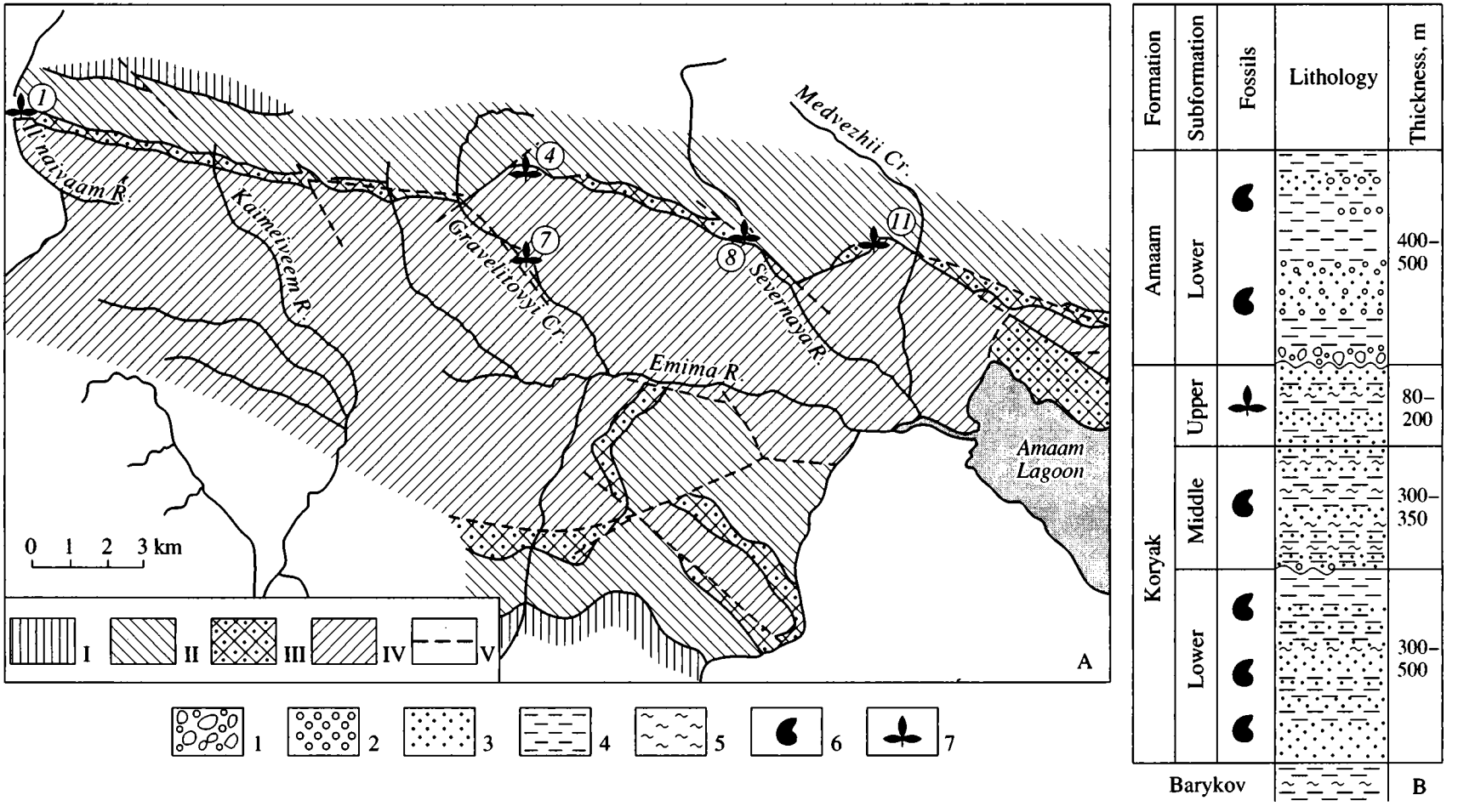

Fig. 2. Schematic geological map (A) of the Emima–Il'naivaam Interfluve, the Amaam Lagoon area (after Volobueva and Terekhova with additions), and lithostratigraphic column (B): (I) Barykov Formation, (II) lower and middle subformations of the Koryak Formation, (III) upper subformation of the Koryak Formation, (IV) Amaam Formation, (V) fault, encircled figures denote fossil flora localities; (1) conglomerate, (2) gritstone, (3) sandstone, (4) clay siltstone, (5) tuff, (6) fossil fauna, (7) fossil flora.

Taxonomic composition of the Koryak (Northeastern Russia) and Sagwon (North Alaska) taphofloras

Species	Koryak taphoflora	Sagwon taphoflora	Species	Koryak taphoflora	Sagwon taphoflora
Equisetum arcticum Heer	+	+	*Trochodendroides* ex gr. *vassilenkoi* Iljinskaya et Romanova		+
Onoclea hesperia R.W. Brown	+	+	*Trochodendroides* sp.	+	
Coniopteris tschuktschorum (Krysht.) Samylina	+		*Nyssidium arcticum* (Heer) Iljinskaja	+	+
Ginkgo ex gr. *adiantoides* (Ung.) Heer	+		"*Cocculus*" sp.		+
Pseudolarix arctica Krysht.	+		*Platimelis platanoides* Golovn.	+	
Pityophyllum sp.	+		*Rarytkinia terechovae* Vassilevsk. et Golovn.	+	+
Sequoia sp.	+		"*Platanus*" *raynoldsii* Newberry	+	
Metasequoia occidentalis (Newb.) Chaney	+	+	"*Platanus*" sp.		+
Metasequoia sp. (cones)	+	+	*Menispermites* (?) sp.		+
Glyptostrobus nordenskioldii (Heer) R.W. Brown	+		*Corylites beringianus* (Krysht.) Herman	+	+
Mesocyparis beringiana (Golovn.) McIver et Aulenback	+		"*Rubus*" sp.	+	+
Mesocyparis (?) sp.		+	*Celastrinites septentrionalis* (Krysht.) Golovn.	+	
Elatocladus talensis Golovn.	+		*Castaliites inordinatus* Hollick		+
Haemanthophyllum cordatum Golovn.	+		*Quereuxia angulata* (Newberry) Kryshtofovich		+
Haemanthophyllum ex gr. *kamtschaticum* Budantsev		+	*Zizyphus* sp.	+	
Phragmites alaskana Heer		+	*Viburnum rarytkense* Budantsev		+
Phragmites sp.		+	*Cissites pekulneense* (Philipp.) Moiseeva	+	
Sparganiophyllum sp.	+		*Cissites hermanii* Moiseeva	+	
Trochodendroides arctica (Heer) Berry		+	*Liriophyllum* cf. *aeternum* Golovn.	+	
Trochodendroides bidentata Vassilevsk. et Golovn.	+		*Carpolithes biloculatus* Golovn.	+	
Trochodendroides emimensis Golovn.	+		*Dicotylophyllum* sp. 1	+	
Trochodendroides lamutensis Golovn.	+	+	*Dicotylophyllum* sp. 2		+

Emima basin during summer of 1987 and by myself in the Il'naivaam basin in 1990 are stored at the Geological Institute of Russian Academy of Sciences (Moscow), collection no. 4843.

In the Emima basin, the Koryak Formation above the Barykov Formation is divided into three subformations (Herman, 1999). The lower and middle subformations are composed of terrigenous–volcanogenic deposits enclosing a diverse molluscan fauna of Campanian–early Maastrichtian age (Fig. 2). The upper subformation of continental deposits bearing abundant plant remains.

In the lower Koryak Subformation, prevailing volcanomictic to polymictic sandstones are intercalated with subordinate beds of siltstones, tuffs, and tuffstones of acidic composition. Based on abundant ammonites and inocerams found in the unit, it attributed to the lower half of the upper Campanian (Volobueva and Terekhova, 1974). The subformation is 300–500 m thick.

The middle Koryak Subformation (300–350 m) is mainly composed of acidic tuffs, tuffites, tuffstones, and tuffaceous siltstones. Rare *Inoceramus* ex gr. *balticus* Boehm., *Acila* sp. indet., *"Parallelodon" sachalin-*

ensis Schmidt, *Thracia* sp., *Goniomys* sp. indet., *Hypophylloceras* sp. indet., *Gaudryceras* sp., *Canadoceras newberryanum* (Meck), *C. kossmati* Mat., and *C.* aff. *multicostatum* Mat. indicate that this subformation corresponds in range to the terminal upper Campanian–lowermost Maastrichtian (Volobueva and Terekhova, 1974; Volobueva and Krasnyi, 1979). In addition, the early Maastrichtian *Patagiosites alaskensis* Jones and *Inoceramus pilvoensis* Sok. (identified by V.P. Pokhialainen in 1990) were found in the unit (Herman, 1993).

The upper Koryak Subformation bearing plant remains conformably overlies the middle unit. It is composed mainly of sandstones and tuffstones alternating with siltstones, carbonaceous siltstones, rare conglomerate interbeds, and coal lenses. The rocks are 80 to 200 m in total.

The Amaam Formation of conglomerates, gritstones, sandstones, and siltstones rests on a slightly eroded surface of the Koryak Formation. The lower part of the Amaam Formation yields endemic mollusks *Glycymeris amaamensis* Volobueva, *Crassatella* cf. *unioides* (Stanton), *Herella* sp., *Turitella* aff. *pachecoensis* Stanton, the Maastrichtian–Paleocene foraminifers (identified by M.Ya. Serova, see in Volobueva and Terekhova, 1974), and some malacostracans and columnals of the Late Cretaceous crinoid *Wielsenicrinus* (Dundo *et al.*, 1974). *"Parallelodon"* sp. and remains of inoceramid prismatic layers were detected in the lower part of the Chukotka Formation (an equivalent of the Amaam Formation) in the Al'katvaam River valley. These fossils indicate the Maastrichtian age of lowermost beds in the Amaam Formation. Thus, the flora-bearing part of the Koryak Formation is within the early–late Maastrichtian interval (basal portion excluded) or in the late Maastrichtian only (Herman, 1999).

We studied both collections of fossil plants sampled from five localities of the upper Koryak Subformation along the northern edge of the Amaam depression (Fig. 2). Plant remains occur in lower and upper parts of the subformation, and sampled exposures are spaced apart for about 25 km at maximum. Because of persistent taxonomic composition throughout the subformation section, plant remains are considered as representing a single taphoflora. The Koryak taphoflora includes 31 species (table) of Equisetites, ferns, ginkgophytes, conifers, and angiosperms.

Rare *Equisetum arcticum* found in the localities is the only taxon of Equisetites. Innumerous ferns are represented by sparse *Onoclea hesperia*, a characteristic component of the Koryak flora, and by fragments of *Coniopteris tschuktschorum.* Sporadic species *Ginkgo* ex gr. *adiantoides* represent ginkgophytes.

Conifers and angiosperms are dominants of a very diverse composition. Prevailing among conifers are species *Metasequoia occidentalis* and *Glyptostrobus nordernskioldii* of Taxodiaceae and *Mesocyparis beringiana* cupressaceous. They occur in association with *Metasequoia* cones and sparse impressions of morphologically peculiar shoots (lacking cuticle) and broad flattened *Sequoia* leaves. Pinaceae are represented by rare impressions of long *Pityophyllum* needles and fruit scales of *Pseudolarix arctica. Elatocladus talensis* having shoots with dense linear large leaves arranged spirally was identified among conifers of unclear systematic position.

Angiosperms, primarily arboreal dicotyledons are also quite abundant and diverse, being dominated by *Corylites beringianus, "Platanus" raynoldsii*, and *Trochodendroides.* The last genus is represented by four species: *Trochodendroides bidentata* with large crenate and bicrenate leaves, less frequent *T. lamutensis* and *T. emimemsis*, and by leaves with pinnate venation of *Trochodendroides* sp., which may correspond to first leaves of shoots. Fruits and leaves of *Nyssidium arcticum* occur associated.

Characteristic but infrequently dominant species of the Koryak flora are *Cissites pekulneense, C. hermanii, Liriophyllum* cf. *aeternum, Platimelis platanoides, Celastrinites septentrionalis*, and *Rarytkinia terechovae.* They are typical the Koryak flora and the Rarytkin plant assemblage from the Anadyr' River basin but have no affinity to other floras of the Northern Hemisphere. The flora under consideration also includes single leaf imprints of *"Rubus"* sp., *Zizyphus* sp., *Dicotylophyllum* sp. 1, and small fruits of *Carpolithes biloculatus.* Monocots are represented by semiaquatic *Haemanthophyllum cordatum* and by oblong linear leaves of presumably aquatic herb *Sparganiophyllum* sp.

The Koryak flora has a little in common with younger Cretaceous floras of the Ugol'naya Bay and Rarytkin Range. The Campanian Barykov flora of the Ugol'naya Bay (Herman, 1999) is dominated by *"Quercus" tchucotica, Macclintockia ochotica, Macclintockia beringiana*, and by cycadophytes. *Paraprotophyllum ignatianim* is most abundant among Platanaceae. These taxa are absent in the Koryak flora. Taxa of a wide stratigraphic range, such as *Equisetum arcticum, Ginkgo* ex gr. *adiantoides, Metasequoia occidentalis, Glyptostrobus nordenskioldii*, and *Nyssidium arcticum*, represent species in common for the Koryak flora and early Maastrichtian Gornaya Rechka assemblage of the Anadyr' basin (Golovneva, 1994). *Mesocyparis beringiana, Corylites beringianus*, and *Celastrinites septentrionalis* are scarce in the above assemblage, being abundant in the Koryak flora.

In its general aspect, the Koryak flora is similar to the Danian–early Paleocene floras of Northeastern Russia, especially to the Rarytkin flora of the Anadyr' basin (Golovneva, 1994). In these floras, Equisetales, ferns, and ginkgophytes are relatively scarce and conifers are less abundant than angiosperms plants and include dominant *Metasequoia occidentalis, Mesocyparis beringiana*, and *Glyptostrobus nordenskioldii.* The Koryak and comparable floras of northeastern Russia contain *Elatocladus talensis.* Taxa prevailing among

angiosperms are *Corylites beringianus*, several *Trochodendroides* species (first of all, *T. bidentata*), and *"Platanus" raynoldsii. Cissites pekulneense, Platimelis platanoides, Celastrinites septentrionalis, Rarytkinia terechovae, Liriophyllum aeternum*, and *Haemanthophyllum cordatum* are species in common.

The Koryak flora also shows a close similarity with other Early Paleogene floras of the Russian Far East. Like the Danian Boshnyakovo assemblage of Sakhalin (Krassilov, 1979), it is dominated by leaves of *Corylites* and by genera *Metasequoia* and *Glyptostrobus* of Taxodiaceae, all associated with *Trochodendroides, Celastrinites*, and *Liriophyllum*. Abundant *Trochodendroides* leaves, *Onoclea, Ginkgo, Metasequoia, Mesocyparis*, and *"Platanus" raynoldsii* remains represent taxa in common for the Koryak flora and Danian Tsagayan assemblage from the Amur area (Krassilov, 1976; Akhmet'ev *et al.*, 2002). Predominance of *Metasequoia, Corylites*, and *Trochodendroides*, as well as presence of *Onoclea, Coniopteris, Equisetum, Ginkgo*, and Cupressaceae are characteristic of both the Malaya Mikhailovka flora from lower courses of Amur River (Akhmet'ev and Golovneva, 1998) and the Koryak flora.

THE SAGWON FLORA OF NORTH ALASKA

Compact coarse-grained sandstones and conglomerates of the Upper Cretaceous–lower Paleocene Prince Creek Formation and of the Paleogene Sagavanirktok Formation constitute the topmost beds of hills from mile 352 to mile 354 of the Dalton highway. The rocks are best exposed in the Sagavanirktok River scarp near the Sagwon Locality, and their outcrops are traceable downstream for a distance of 8 to 10 km (Fig. 3). The younger, upper Paleocene–Eocene non-marine deposits of the Sagavanirktok Formation are well exposed in the Franklin Bluff area, the eastern bank of the Sagavanirktok River, from mile 381 to mile 395 of the Dalton highway (Mull and Harris, 1989).

Abundant plant fossils were collected from the upper part of the Prince Creek Formation (the left bank of the Sagavanirktok River near the Sagwon Locality, Fig. 3) by J.A. Wolfe in 1951 (the unnumbered collection of USGS is stored in the Smithsonian Institute, Washington, USA). Spicer collected here plant remains in 1987 and Spicer, Ahlberg, and myself in 2001 (collections of USGS 11481, 87RAS, and 01RAS are in the Open University, Milton Keynes, Great Britain). In river scarps confined to axial zone and to northern and southern wings of the Ivishak anticline, there are exposed terrigenous and coal-bearing deposits of the upper Prince Creek Formation of the Late Cretaceous–early Paleocene age, as well as sandstones and conglomerates of the basal Sagwon Member of the Sagavanirktok Formation of the late Paleocene–Miocene age (Mull and Harris, 1989; Mull *et al.*, 2003).

In above areas, the upper part of the Prince Creek Formation is composed of alternating conglomerates, poorly cemented sandstones, siltstones, carbonaceous shales, and coals, the latter of autochthonous (carbonized wood) and, probably, of allochthonous types. Three coal layers reaching 7 m thickness (the upper layer) are well traceable in all exposures being the marker horizons. Plant fossils were found mainly in the siltstones, carbonaceous and clayey shales, and in siderite concretions, hosting the best-preserved remains. The flora-bearing beds of the Sagwon Locality are 100 to 115 m thick in general.

Basal 25- to 30-m-thick beds of the Sagavanirktok Formation are represented by light colored and whitish, poorly cemented quartz sandstones white at the weathered surface, and by sands, conglomerates, and gritstones. The rocks are exposed in river scarps downstream from the Sagwon Locality (Mull and Harris, 1989; Mull *et al.*, 2003). Their lower unexposed contact appears to be sharp, corresponding to a certain stratigraphic unconformity (Mull *et al.*, 2003). The upper, poorly exposed part of the Sagwon Member is composed of poorly lithified clayey and coaly deposits. The overall thickness of the Sagwon Member of the Sagavanirktok Formation is up to 900 m.

Flora-bearing coaliferous deposits of the Sagwon Locality were previously attributed to the lower Sagavanirktok Formation (the Sagwon Member), and the Cretaceous–Paleogene boundary was positioned within them (Spicer *et al.*, 1994). Later on it was proposed to distinguish this boundary between the Prince Creek and Sagavanirktok formations above the coal-bearing unit and at the base of well traceable whitish sandstones, conglomerates, and gritstones, because that unit is similar in lithology to underlying rocks rather than to overlying strata (Mull *et al.*, 2003).

Flora-bearing deposits of the upper Prince Creek Formation were dated back to the early Paleocene based on palynological data (Frederiksen *et al.*, 1996; Mull *et al.*, 2003). The oldest, coal-bearing deposits of the Sagwon Locality (sections B1 and B2), contain several pollen taxa typical of the Paleocene along with characteristic Late Cretaceous–Early Paleogene forms (Frederiksen *et al.*, 1996). As stated in the mentioned work, prevalence of *Triporopollenites mullensis* indicates the Paleocene or, more likely, the early Paleocene age of host deposits. Stratigraphically higher, the upper, coaly part of the Sagwon Member (section B3 downstream the Sagavanirktok River, at the Ivishak River mouth) yielded a palynomorph assemblage presumably of the late Paleocene age (Frederiksen *et al.*, 1996). Accordingly, the Sagwon Member (and, subsequently, the basal part of the Sagavanirktok Formation) is the late Paleocene in age (Mull *et al.*, 2003).

We studied all four collections of plant fossils from the upper Prince Creek Formation of the Sagwon locality, the eponymous site used to name the flora. Plant remains have been collected from several outcrops on

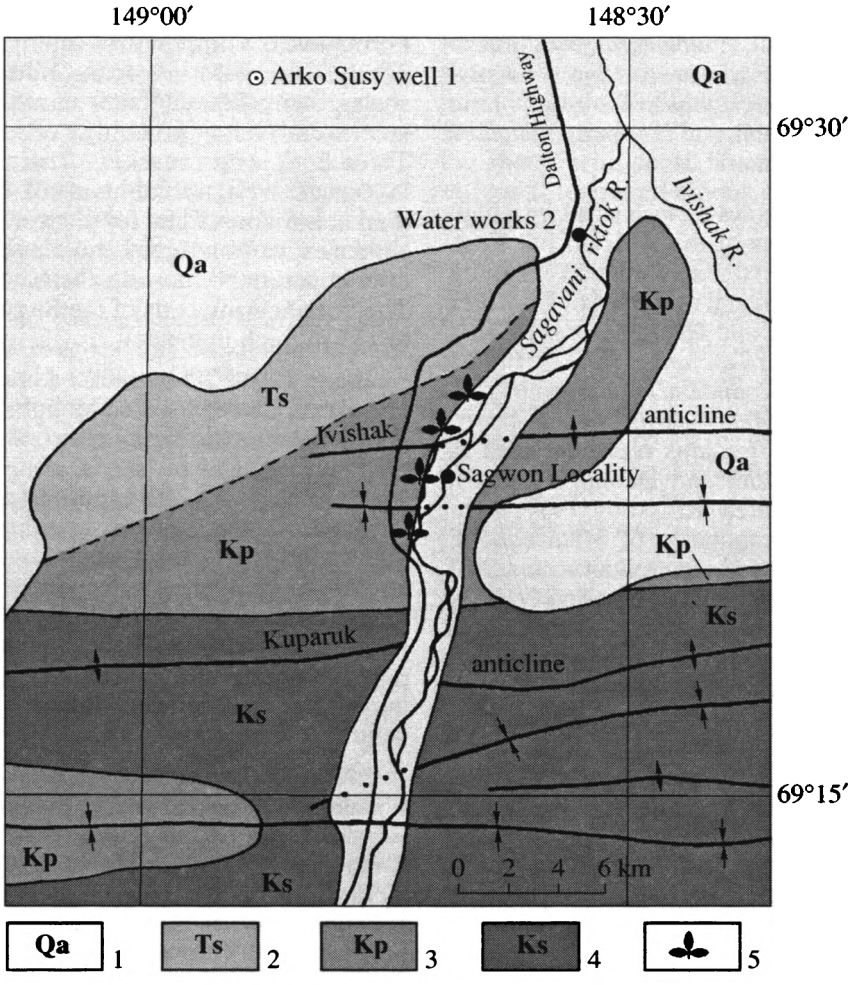

Fig. 3. Schematic geological map of the Sagavanirktok River basin in the Sagwon Locality (after Mull and Harris, 1989, with modifications): (1) undifferentiated alluvial deposits, Quaternary; (2) Sagavanirktok Formation, Upper Paleocene; (3–4) Colville Group: (3) Prince Creek Formation, Upper Cretaceous-lower Paleocene; (4) Schrader Bluff and Seabee formations, Upper Cretaceous, (5) fossil locality.

the left bank of the Sagavanirktok River (Fig. 3) and united into a single assemblage owing to a great taxonomic similarity. The Sagwon taphoflora includes 22 species (table), fructifications of two types (most likely of angiosperms), an imprint of a long shoot with short attached branches, and an imprint of presumably aquatic plant stem.

In general, the Sagwon flora is of a relatively low taxonomic diversity. It is interesting that the early Paleocene palynological assemblages from the same deposits are also of low diversity, which gradually increased in North Alaska throughout the Paleocene and Eocene (Frederiksen *et al.*, 1996, 1998).

Equisetites are represented by *Equisetum arcticum* and ferns by *Onoclea Hesperia.* Both species are rare. Conifers of low abundance and diversity dominate in few localities. Their most frequent components are *Metasequoia occidentalis* shoots with leaves, whereas fragmented specimens of *Mesocyparis* are rare. Monocots are represented by probably semiaquatic taxon *Haemanthophyllum* ex gr. *kamtschaticum.* Oblong leaves with parallel venation, which are traditionally referred to the genus *Phragmites*, may belong to aquatic monocots. These plants together with *Quereuxia angulata* constitute "an aquatic community" of the Sagwon taphoflora, which grew in swampy oxbow lakes.

The greater part of angiosperm leaves belonged to arboreal and bushy dicots. Polymorphous and frequently large-sized leaves of *Corylites beringianus* prevail among them. It is to be noted that palynomorph assemblages from flora-bearing deposits are dominated by anemophilous (wind-pollinated) triporate species *Paraalnipollenites alterniporus* and *Triporopollenites mullensis*, which, according to Frederiksen *et al.* (1996), might be produced by plants with the *Corylites beringianus* leaves. Some localities contain abundant *Trochodendroides* leaves (three species) in addition to

Nyssidium and *Quereuxia* leaves associated with fruits. Remains of other angiosperms are infrequent or single.

The Sagwon taphoflora shows the closest similarity to the late Maastrichtian Koryak flora of the Amaam Lagoon of Northeastern Russia. These floras are compared in detail below. Another taxonomically similar fossil flora is probably the late Maastrichtian flora from the upper half of the Rarytkin Formation of the eponymous ridge (Golovneva, 1994). It includes almost all plant species found in the Sagwon taphoflora.

The Sagwon flora is also similar to some Early Paleogene floras of the Russian Far East. The early Danian Boshnyakovo flora of Sakhalin is also dominated by *Metasequoia, Corylites, Platanus*, and *Trochodendroides*, although it includes cycadophyte *Nilssonia* in distinction (Krassilov, 1979). The late Maastrichtian–early Danian flora of the Yurii Island in the Lesser Kurile group includes abundant Taxodiaceae and Cupressaceae along with angiosperm genus *Corylites* (Krassilov *et al.*, 1988). Abundant remains of *Metasequoia, Trochodendroides* ex gr. *arctica, Corylites*, and *Equisetum arcticum, Onoclea, Woodwardia, Glyptostrobus*, and others are known in the Danian Malaya Mikhailovka flora from lower courses of the Amur River (Vakhrameev and Akhmet'ev, 1977). *Metasequoia, Thuja, Trochodendroides arctica, "Corylus,"* and other representatives of Betulaceae dominate in the early Paleocene Takhobe flora of Sikhote-Alin' (Vakhrameev and Akhmet'ev, 1977).

Thus, most characteristic taxa of the Sagwon flora (*Equisetum arcticum, Onoclea, Metasequoia, Mesocyparis* (?) cf. *Rubus, Trochodendroides, Corylites*, and probably *"Platanus" raynoldsii*) are also typical of the Danian–early Paleocene floras of the Far East and Northeastern Asia. The Sagwon flora characterizes typical vegetation of deciduous broad-leaved and mixed coniferous–broad-leaved forests, which grew in wide river valleys with swampy oxbow lakes of subarctic zone under temperate (probably warm–temperate) and humid climate conditions.

FLORAL CHANGES ACROSS THE CRETACEOUS–PALEOGENE BOUNDARY IN THE NORTH PACIFIC

The described Koryak taphoflora of Northeastern Russia and the Sagwon taphoflora of North Alaska are suitable for investigation of taxonomic changes in floras of the North Pacific across the Cretaceous–Paleogene boundary that marks one of most significant ecological crises of the Earth history (Fig. 4). The late Maastrichtian age of the Koryak flora is reliably established based on correlation of flora-bearing continental and marine deposits, because the latter yield the age-diagnostic faunal remains. The Sagwon taphoflora is dated as the early Paleocene in age, and this assessment based on palynological data but not on marine fossils appears to be sufficiently reliable as well. The Rarytkin flora from the eponymous ridge (Golovneva, 1994) is also of interest, because the Cretaceous–Paleogene boundary is suggested to be within its stratigraphic range that is established, though less reliably, based on plant macrofossils (Fig. 4).

The Koryak taphoflora has much in common with the Sagwon taphoflora. Dominant components in both floras are angiosperm dicots, whereas content of Equisetites forms and ferns is lower. Conifers rather diverse and abundant in the Koryak flora were important components of vegetation cover. The Sagwon flora of a lower taxonomic diversity includes nevertheless almost all the components established in the Koryak flora (table). *Corylites beringianus*, several *Trochodendroides* species, *Nyssidium* leaves with associated fruits, and *Metasequoia occidentalis* are dominants of both floras. The other taxa in common are *Equisetum arcticum, Onoclea hesperia, Rarytkinia terechovae, Haemanthophyllum*, and *"Rubus"*.

At the same time, the Sagwon flora contains fewer conifers than the Koryak flora. It is devoid of *Glyptostrobus, Elatocladus, Pityophyllum*, and *Pseudolarix*, while remains of *Mesocyparis* are scarcer and fragmentary. The Sagwon flora is also lacking the characteristic Koryak dicots *Cissites, Celastrinites, Platimelis*, and *Liriophyllum*, whereas *"Platanus"* is represented by single specimens. Absence of characteristic Cretaceous taxa indicates a younger age of the Sagwon taphoflora that is consistent with palynological data implying the early Paleocene age of this flora (Frederiksen *et al.*, 1996). *Castaliites inordinatus, Quereuxia angulata*, and *Viburnum rarytkense* are the unique taxa of the Sagwon flora.

The fact that the late Maastrichtian Koryak flora of the Amaam Lagoon site is very similar in taxonomic composition and general aspect to the early Paleocene Sagwon flora of northeastern Alaska (Fig. 4) evidences against a hypothesis of a global ecological crisis at the Cretaceous–Paleogene boundary (Spicer, 1989; Spicer *et al.*, 1994). In any case, the crisis did not result in a considerable extinction of plants and had no significant effect on evolution of the North Pacific floras.

Golovneva (1990, 1994) who studied the Rarytkin flora arrived at the same conclusion. In her opinion, the Cretaceous–Paleogene boundary is within the flora-bearing beds of the Rarytkin Formation at the level separating the early and late subassemblages of the flora (Fig. 4). If this is correct, then it is very significant that floral changes across the boundary are unimportant, allowing the floras of each side of the boundary to be regarded as subassemblages of the single Rarytkin taphoflora. "The subassemblages of the Rarytkin Formation flora are not sharply different from each other" (Golovneva, 1994, p. 41). More essential taxonomic changes in plant assemblages are recorded in the Maastrichtian part of the Rarytkin Formation (approximately in the middle or upper Maastrichtian), where the Rarytkin assemblage replaces the Gornaya Rechka flora

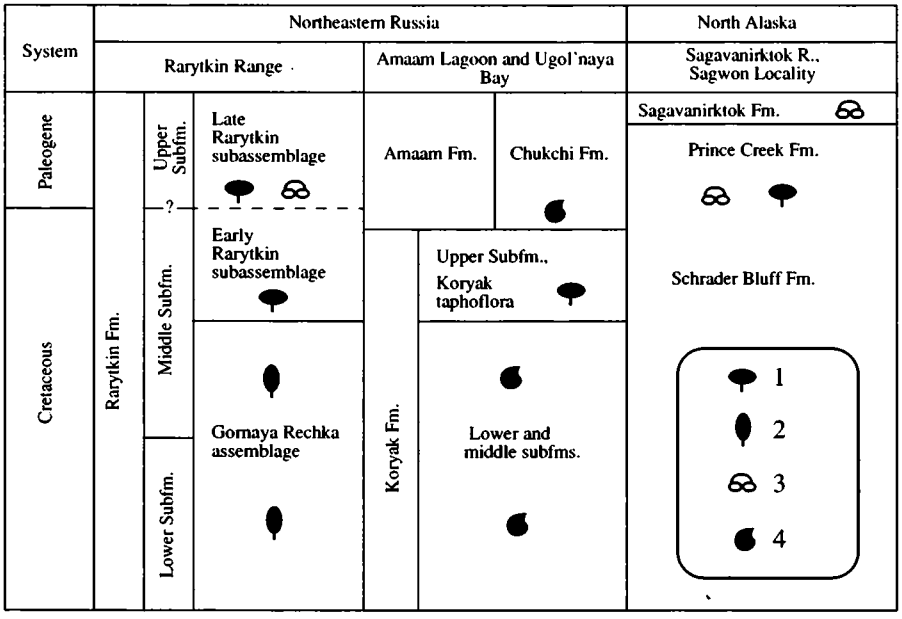

Fig. 4. Floral changes in North Alaska and northeastern Russia at the Cretaceous–Paleogene boundary: (1) taphofloras of the Koryak and Sagwon evolutionary stages, (2) taphofloras of the Gornaya Rechka stage; (3) age-controlling palynological assemblages; (4) age-controlling marine mollusks.

(Fig. 4). Golovneva (1994) noted that the Rarytkin taphoflora was distinguished from the Gornaya Rechka assemblage based on several features. These were "...a noticeable change in species composition of ferns, gymnosperms, and angiosperms, an alteration of dominant species, the replacement of older conifer taxa (*Sequoia minuta, Crypromerites, Parataxodium*) by younger ones (*Metasequoia, Glyptostrobus, Taxodium*), and diminishing proportion of extinct form genera versus growing abundance of recent genera" (Golovneva,1994, p. 41). As is concluded in her work, floras of Danian affinity substituted the Maastrichtian type flora in the Koryak Upland at the middle–late Maastrichtian boundary time and radiated then southward during the progressing cooling.

Paleobotanic evidence for catastrophic events recorded at the Cretaceous–Paleogene boundary are known mostly from middle-latitude North America (Spicer, 1989; Spicer *et al.*, 1994). Data considered above imply however that the most significant (but not catastrophic!) change in the North Pacific floras took place in the Maastrichtian time rather than at the Cretaceous–Paleogene boundary (Fig. 4). Changes across the latter are no larger in scale than those known in the "crises-free" Late Cretaceous (Golovneva, 1990, 1994; Golovneva and Herman, 1998; Spicer *et al.*, 1994; Herman and Spicer, 1995, 1997). Thus, the high-latitude floras of Asia and North America (Johnson, 1988; Johnson *et al.*, 1989) do not evidence in favor of catastrophic event at the Cretaceous–Paleogene boundary time. On the contrary, development of floras was likely controlled here by long-term climatic fluctuations and by evolution and migration of plants. As the essential changes in composition of Late Cretaceous floras are untypical of the North Pacific region, it is possible to conclude that presumable consequences of meteoritic impacts (mass extinction of plants as a result of cooling and sharply depressed photosynthesis in response to reduced transparency of the atmosphere) are recorded in the middle-latitude North America only and have not been global in scale.

A similar development of floral changes across the Cretaceous–Paleogene boundary are recognized in the southeastern part of the Zeya–Bureya basin (Akhmet'ev *et al.*, 2002). Flora from the upper part of the middle Tsagayan Subformation is of the Late Maastrichtian age, as is established based on palynomorph assemblages, plant macrofossils (data of Markevich and Bugdaeva in *Flora...*, 2001), and some other geological records. The Danian (early Danian?) flora of the upper Tsagayan Formation includes many plant remains characteristic of the middle subformation and associated with some typical Paleogene taxa. "Taxonomic analysis of floral assemblages from the middle and upper subformations shows that they differ from each other mainly due to appearance of new species" (Akhmet'ev *et al.*, 2002, p. 313). In other words, evidence in favor of catastrophic events at the Cretaceous–Paleogene boundary, i.e., any noticeable extinction of plants, has not been discovered here as well.

The Maastrichtian–Eocene palynological assemblages of North Alaska demonstrate a gradual increase in palynomorph diversity from low in the early Paleocene to very high in the early and middle Eocene, and this is thought to be an indication of floral restoration after mass extinction in the terminal Cretaceous (Fred-

eriksen *et al.*, 1996, 1998). However, another interpretation is also possible. The gradual increase of taxonomic diversity in the North Alaskan floras could be a consequence of progressing warming from the pessimum phase in the second half of the Maastrichtian, which is inferred from a low diversity of the Kogosukruk flora (Spicer and Parrish, 1990), to the global temperature optimum in the late Paleocene–early Eocene.

As noted above, the Koryak flora is similar in taxonomic composition and general aspect to the Early Paleogene floras of northeastern Asia and North America rather than to the Late Cretaceous floras. The Late Maastrichtian flora of the Koryak Upland might be a source community for many Sagwon plants, species of which migrated to North Alaska from northeastern Asia during the Paleocene warming. Probably, it was the late Maastrichtian Koryak flora that gave rise to the Early Tertiary deciduous floras of the Arctic and adjacent regions (Herman and Spicer, 1995, 1997).

ACKNOWLEDGMENTS

The work was supported by the Federal Scientific School Program, project no. NSh-1615.2003.5, and by the project "Factors of the biosphere evolution", contract no. 100002-251/OHZ-06/183-181/270603-908.

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Stratigraphy and Geological Correlation. Vol. 12. No. 5. 2004. pp. 495–513. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya. Vol. 12. No. 5. 2004. pp. 65–86.
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New Data on the Marine Paleogene of the Southern West Siberian Plate, Paper 2

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Received December 24, 2002

Abstract—Age and spatial relationships between the Gan'kino, Lyulinvor, and Tavda formations, the principal subdivisions of marine Paleogene, are clarified based on comprehensive paleontological study of marine deposits, which have been recovered by two reference boreholes in the Southern West Siberian plate. As was suggested earlier, the Talitsa Formation is missed from the sequence in a greater part of the Barabinsk lithologic-facies zone. In the study area, accumulation of the Lyulinvor Formation commenced somewhat later than in central and western areas of the plate. Breaks in sedimentation are established inside and at the top of the Lyulinvor Formation below the Tavda Formation base. Sedimentation rates are evaluated, and lithological and biostratigraphic subdivisions are correlated with geological and paleogeographic events. Principal transgressive–regressive cycles and half-cycles are outlined. It is shown for the first time that accumulation of the Tavda Formation corresponded to two cycles of this kind, the second of which is represented solely by its transgressive part corresponding to Lattorfian (late Priabonian) eustatic transgression. Directions of block movements, which controlled either the basin free connection with or partial isolation from ocean, are established. Impacts of hydrology and principal Arctic and Tethyan currents on sedimentation and climate are considered.

Key words: Paleogene, Eocene, Oligocene, West Siberian plate, foraminifers, dinocysts, radiolarians, diatoms, ostracodes, paleomagnetic zones, magnetostratigraphy, Gan'kino, Lyulinvor, and Tavda formations.

INTRODUCTION

The first our paper (Akhmet'ev *et al.*, 2004) was devoted to biostratigraphic characterization of two reference boreholes, which recovered the marine Paleogene deposits of the Southern West Siberian plate. In that paper, we presented data on the sequence structure and distribution of various fossil groups (planktonic and benthic foraminifers, radiolarians, diatoms, organic-walled phytoplankton, spores and pollen) in lithostratigraphic units. Magnetostratigraphic and logging results were presented as well. In this work, we discuss the results obtained.

AGE SUBSTANTIATION

Gan'kino Formation

The Gan'kino Formation age is defined based on its nannoplankton assemblage that is characteristic of the *Nephrolithus frequens* Zone (zone CC26 corresponding to the second half of late Maastrichtian in the International standard zonation of the Upper Cretaceous). The assemblage diversity suggests a warm-water state of the Gan'kino sea basin that was of normal salinity and connected with the World Ocean. The inference is consistent with data on ostracodes and benthic foraminifers. The assemblage of the latter can be attributed to the *Spiroplectammina carinata* Zone of the upper Maastrichtian.

Earlier N.K. Lebedeva and A.F. Khlonova studied Maastrichtian dinocysts of West Siberia in the Ust-Yenisei depression only (Zakharov *et al.*, 1986; Il'ina *et al.*, 1994). Unfortunately, the host deposits have been dated then in the stage rank only, because the assemblages studied were taxonomically incomplete. Specifying age of the Gan'kino assemblage, Aleksandrova considered available data on Greenland, North Sea, West Europe, and Mangyshlak (Sharafutdinova, 1992;

Hansen, 1977, 1979; Hakånson *et al.*, 1979; Kirsh, 1991; Marheinecke, 1992; Schøler *et al.*, 1992; Brinkhuis *et al.*, 1996; Herngreen *et al.*, 1998; Slimani, 2001), as the dinocysts zonation in West Siberia is known inadequately. In the North Sea, Netherlands, and Germany, assemblages of most close composition are characteristic of the upper Maastrichtian. Based on dinocyst species in common, such as *?Phanerodinium sonciniae, P. veligerum, Cordosphaeridium fibrospinosum, Conneximura fimbriata, Cerodinium speciosum, Cerodinium diebelii, Hystrichostrogyton coninkii*, and *Trythirodinium evittii*, she identified the mid-late Maastrichtian age of the Gan'kino Formation, because *Palynodinium grallator* (the index species of the terminal Maastrichtian), *Thalassiphora pelagica* and Paleogene taxa are missed from the studied assemblage. The *Cerodinium diebelii* Beds of the formation are correlative therewith to the *Isabelidinium cooksoniae, Palaeocystodinium denticulatum*, and *Hystrichostrogylon borisii* zones of the North Sea (Schøler *et al.*, 1992), and to *Deflandrea galeata* Zone of Germany (Kirsch, 1991), Belgium (Slimani, 2001) and Dania (Fig. 2).

Two magnetic zones of normal and reversed polarity established in the Gan'kino Formation clays are of insignificant thickness and can be correlated with the boundary interval of terminal C30n and basal C29r chrons (Berggren *et al.*, 1995).

Lyulinvor Formation

Lower Lyulinvor Subformation

Characteristic of the lower Lyulinvor Subformation are the *Alisocysta margarita* and *Apectodinium homomorphum* dinocyst assemblages of the Thanetian age. The former assemblage is established between the first occurrence levels of *Deflandrea denticulate* and *Apectodinium homomorphum*, because *Alisocysta margarita* is present, though as single specimens, in the late Selandian sediments. The *A. margarita* Zone is reliably correlated with the West European biostratigraphic subdivisions Viborg4 (Heilmann-Clausen, 1985), partly with D4 (Costa *et al.*, 1988), and with *Alisocysta margarita* Zone of England (Powell, 1992). In the work by Powell, the zone is correlated with the nannoplankton zone NP8 of the lower Thanetian. In Borehole 9, the *Apectodinium homomorphum* Zone is distinguished above the first occurrence level of its index species. In West European sections, the stratotype included, it is correlated with the dinocyst zones Viborg 5 (Heilmann-Clausen, 1985), D4 (Costa *et al.*, 1988), and *Apectodinium homomorphum* of southern England (Powell, 1992), corresponding there to the nannoplankton zone NP9 of the upper Thanetian.

An impoverished assemblage of agglutinated foraminifers from basal beds of the subformation belongs to the regional *Glomospira gordialiformis* Zone. Podobina (1998) attributed this assemblage characteristic of the Talitsa–Lyulinvor transitional strata to the late Paleocene (Thanetian).

Four paleomagnetic zones of normal and reversed polarity (N_1ll, R_1ll, N_2ll, R_2ll), which are detected in the Thanetian interval, may correspond to chrons C26n, C25r, C25n, and C24r or, possibly, to their fragments. This does not exclude, however, that the subformation lower boundary is close to the Selandian–Thanetian transition. Biotic changes across the lower and upper boundaries of the lower Lyulinvor Subformation indirectly suggest breaks in sedimentation at both levels.

Upper Lyulinvor Subformation

According to distribution pattern of siliceous sediments in the section of Borehole 9, the upper Lyulinvor Subformation is divided in two members of (a) gray diatomaceous and opoka-like shales and of (b) green mudstones.

(a) **Member of gray diatomaceous and opoka-like shales** spans the interval of the *Dracodinium varielongitudum* and *Charlesdowniea coleothrypta* s. l. dinocyst zones. In many West European stratigraphic charts, the former zone is in the middle of the Ypresian Stage. Being at the same level in the type section of Ypresian clays in Belgium, the zone is confidently correlated, based on assemblages of calcareous microplankton, with the lower part of the nannoplankton zone NP12 and with the zone P7 in zonation of planktonic foraminifers. However, Andreeva-Grigorovich (1991) correlated the *D. varielongitudum* Zone with an upper part of zone NP11. In the section of Borehole 9, the dinocyst zone in question can be directly correlated with the *Heliodiscus inca* Zone of radiolarian standard, although its basal (491–489 m) and topmost beds belong to the *Heliodiscus lentis* s. str. and *Buriella longa–B. clinata* zones, respectively. According to distribution of diatoms, the upper *Dracodinium varielongitudum* Beds are likely correlative with a lower part of the *Pyxilla gracilis* s. l. Zone of the middle Ypresian. A lower interval of the beds, which is barren of diatoms, seems to be corresponding to the upper *Coscinodiscus payeri* Beds. In sections of boreholes 011 BP, upper strata of the *Dracodinium varielongitudum* Zone yield radiolarians occurring in the terminal part of the *Heliodiscus inca* Zone.

Based on benthic foraminifers occurring in the *Dracodinium varielongitudum* Zone, basal strata of the latter can be correlated with the *Bolivinopsis spectabilis* Beds, whereas its upper part is correlative with the lower interval of the *Textularia carinatiformis* Zone. Planktonic foraminifers *Pseudohastigerina wilcoxensis* and *Subbotina eocaenica* from the depth levels of 477 and 451 m suggest a broad interval from the early to initial middle Eocene for their host deposits. The *D. varielongitudum* Zone spans paleomagnetic zones N_3ll, R_3ll, N_4ll, and R_4ll. Paleomagnetic zone R_1ll distinguished in Borehole 011BP near the top of the *D. vari-*

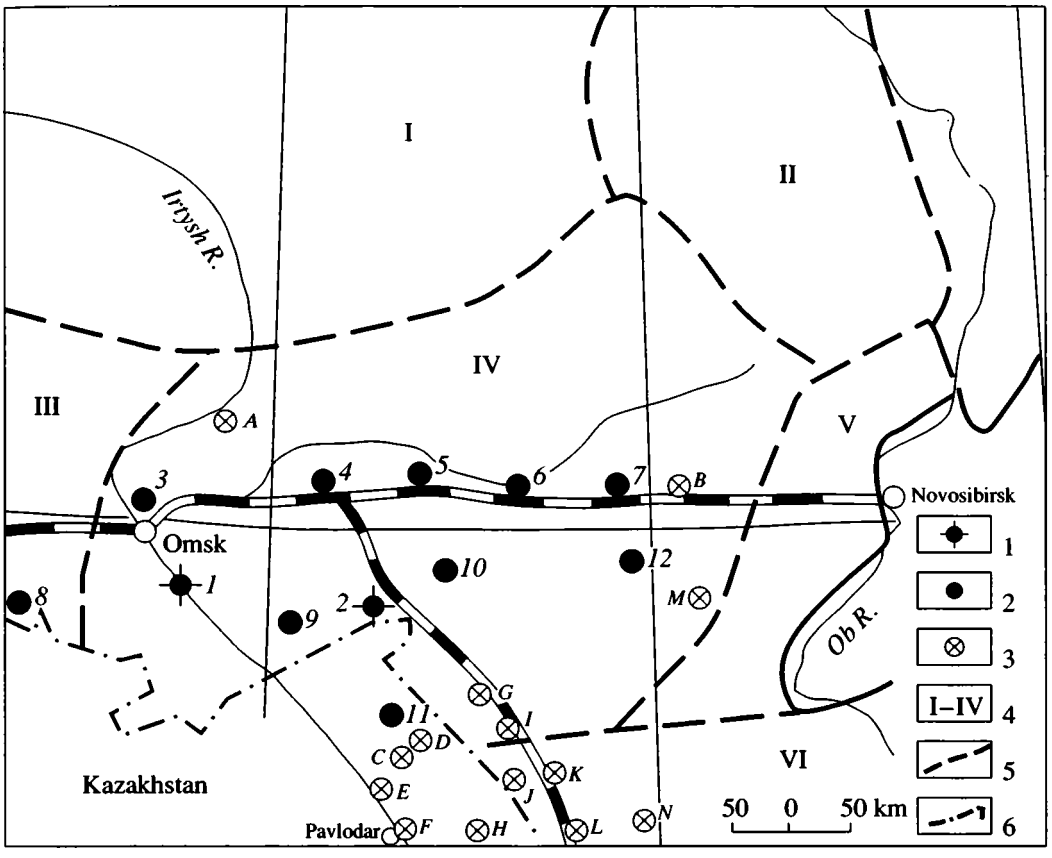

Fig. 1. Lithologic-facies zones and borehole localities discussed in this work (scheme is compiled based on data from *The Unified...*, 2001; Martynov, 1967, 1968, 1973; Zal'tsman, 1968): (1) boreholes 011BP of the Achair area (*1*) and 9 of the Chistoozerninskoe area (*2*) described in this work; (2) Omsk reference borehole (*3*), exploration boreholes 1R and 2R of the Tatarskoe area (*4*), exploration boreholes 1R and 1K of the Tibes area (*5*), Barabinsk reference borehole (*6*), Ubinsk borehole (*7*), Kishikaroi borehole (*8*), boreholes 69 and 2R near the village of Cherlak (*9*), exploration boreholes of the Ipatovo area (*10*), Mikhailovskaya-32 (*11*) and Yarkinskaya-35 (*12*); data on which are summarized in Table 2; (3) boreholes Bol'sherechenskaya (*A*), Kargatskaya-!2 (*B*), Trofimovskaya-30 (*C*), Trofimovskaya-3 (*D*), Kachirskaya-38 (*E*), Sychevskaya-34 (*F*), Karasukskaya 1R and 117 (*G*), Efremovskaya 185 and 48 (*H*), Burlinskaya-5K and others (*I*), Uspenskaya-35 (*J*), Slavgorodskaya-129 and others (*K*), Kulundinskaya-14 (*L*), Kochkinskaya (*M*), Blagoveshchenskaya 1R and others (*N*) discussed in brief; (4) Central (I), Narym (II), Ishim (III), Barabinsk (IV), Priobskaya (V) and Kulunda (VI) lithologic-facies zones; (5) boundaries between zones; (6) Kazakhstan state boundary.

elongitudum Zone is correlative with magnetozone R_4II of Borehole 9. As the *D. varielongitudum* Zone corresponds to the middle Ypresian, the above paleomagnetic zones can be correlated with chrons C24n, C23r, C23n and with the lowermost interval of chron C22r. Rocks with normal polarity prevail in the indicated magnetostratigraphic interval of the upper Lyulinvor Subformation.

In both sections, the well-developed *Charlesdowniea coleothrypta* s.l. Zone of dinocysts is of the late Ypresian–early Lutetian age according to opinion of many researchers. Andreeva-Grigorovich (1991) who studied organic-walled phytoplankton and nannoplankton in southern sections of the former USSR correlated this biostratigraphic unit with standard zones NP12–NP14 and NP15 (partly). Being of the same range in regional stratigraphic charts of France and the North Sea, the unit appears to be at the level of zones P9 and P10 in zonation of planktonic foraminifers and spans zones NP15 (partly) and NP14 of the nannoplankton scale. The *Ch. Aoleothrypta* s. str. and *Ch. Aoleothrypta rotundata* subzones of the *Charlesdowniea coleothrypta* s. l. Zone correspond, respectively, to the Ypresian and terminal Ypresian–initial Lutetian intervals. In the studied sections, the latter subzone is distinguished conventionally because of unclear boundaries. It corresponds to the upper submember of gray diatomaceous clay that is 2 to 5 m thick.

The lower *Charlesdowniea coleothrypta* s.l. beds belong to the uppermost interval of the *Heliodiscus inca* Zone of the radiolarian scale, while the main part of this dinocyst zone is equivalent in range to the *Buriella clinata–B. longa* interval (though these index species have not been encountered in our samples). In Borehole 011BP, the uppermost *Ch. coleothrypta* beds yield radiolarians characteristic of interval transitional to the *Lychnocanium separatum* Zone. As for correlation with biostratigraphic units of the diatom scale, the

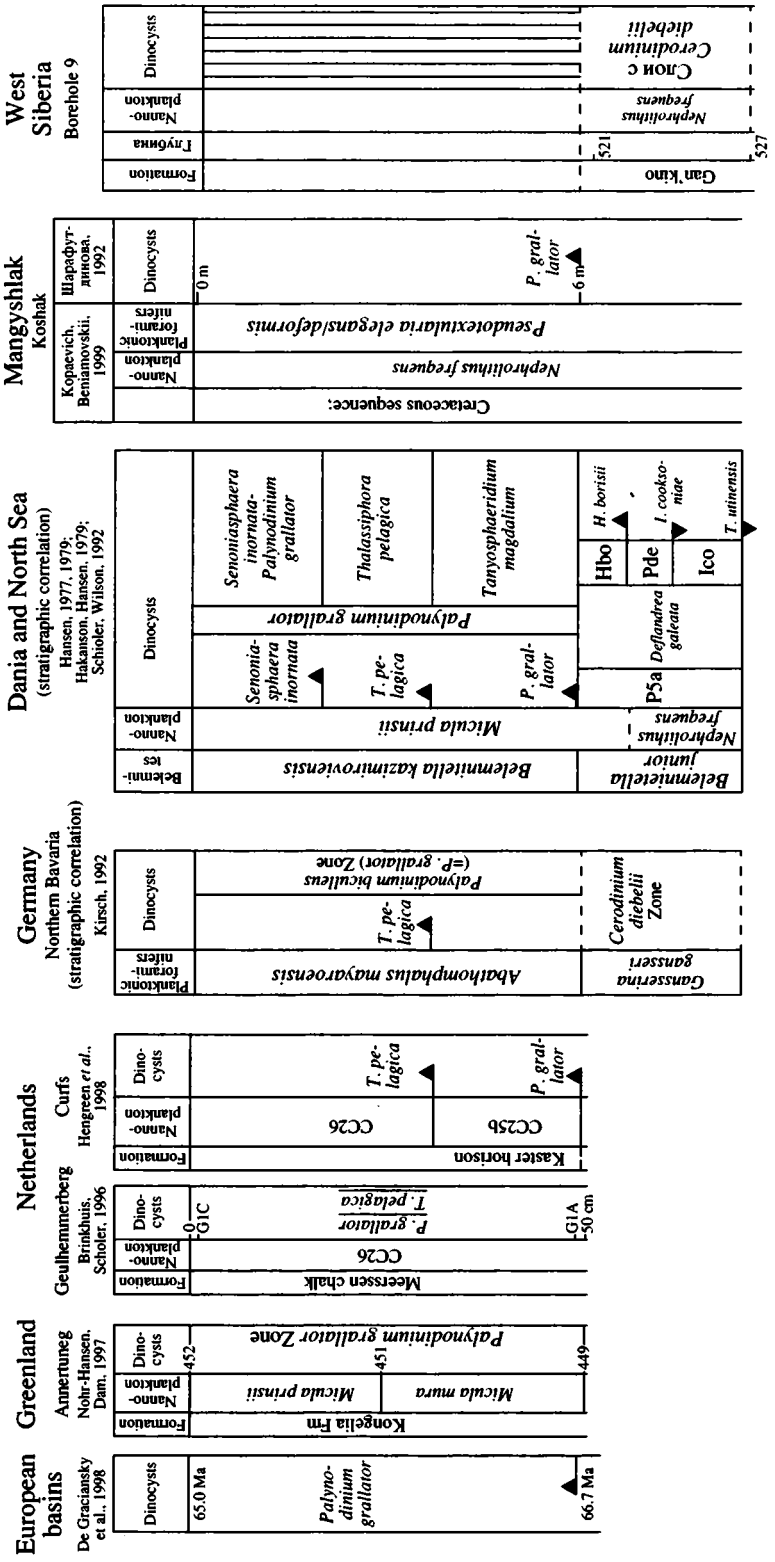

Fig. 2. Correlated upper Maastrichtian dinocyst zonations of different regions.

lower part of the *Charlesdowniea coleothrypta* s.l. Zone is correlative with *Pyxilla gracilis* Zone, while the upper one partially corresponds to the *Coscinodiscus poliactis* Zone. Rubina (1973) who distinguished the last zone considered it as transitional to the *Pyxilla oligocaenica* Zone. In opinion of Strel'nikova, it would be more correct to correlate the upper strata bearing *Ch. coleothrypta* s.l. in Borehole 011BP with the *Pyxilla oligocaenica–P. oligocaenica* var. *tenuis* Zone. Glezer considers the lower *Charlesdowniea coleothrypta* s. str. Subzone as corresponding to the *Brightwellia hyperborean* Zone of diatom scale, which is discriminated from the *Pyxilla gracilis* Zone. In addition, she correlated this subzone with the *Dictyocha secta* Zone of silicoflagellates and suggested the Ypresian age for all the subdivisions. The *Textularia carinatiformis* Zone of agglutinated foraminifers spans the interval of *Charlesdowniea coleothrypta* s.l. Zone coupled with an upper part of older *Dracodinium varielongitudum* Zone.

In Borehole 9, the *Ch. coleothrypta* s.l. Zone ranges through paleomagnetic zones R_4II (upper part), N_5II, and R_5II, which can be correlated with chrons C22r (upper part),C22n, and C21r of the Berggren's scale. According to this correlation, the zone spans the upper Ypresian–lower Lutetian interval. In Borehole 011BP, the dinocyst zone under consideration corresponds to paleomagnetic zones R_1II (upper part) and N_1II correlative with chrons C22r (upper part) and C22n. Rocks of reversed polarity are dominant in this interval of subformation, especially in Borehole 9. Paleomagnetic zones are likely reduced here because of presumable hiatuses characteristic of the interval, which may distort the paleomagnetic records.

(b) **Member of green clays** with fish remains, the terminal one in the upper Lyulinvor Subformation, is most complete in section of Borehole 9, being of a special interest in terms of age determination. The Lyulinvor sedimentation cycle ended with deposition of this member. At this stage of evolution, the West Siberian sea basin under discussion was of a maximum size and occupied about 70% of West Siberia (Shatskii, 1978). Sedimentation rates decreased at that time, although it is difficult to estimate them correctly because of hiatuses distinguished in the section. The reduced thickness of the member definitely reflects a restricted silica influx into the basin, as it follows from sedimentation rates evaluated for lithostratigraphic units of Borehole 9. According to our estimates, consolidated sediments accumulated in a period of 1000 years are 10–12 mm thick in the lower Lyulinvor Subformation, about 10 mm thick in the member of gray opoka-like clay of the upper Lyulinvor Subformation, and only 4 mm thick in the green clay member.

Lower beds of the latter (depth interval of 449.6–444 m) belong to the *Wetzeliella articulata* (acme)–*Systematophora placacantha* Zone of dinocysts. In the biostratigraphic chart of the Parisian basin (Chateaneuf and Gruas-Cavagnetto, 1978), where the Lutetian stratotype was described, *W. articulata* is the index species of zone W9a corresponding to the first half of early Lutetian divided in two substages as in the scheme of Cavellier and Pomerol (1986). The same taxon is indicative of zone D10 (*W. articulata–W. ovalis*) of the late Lutetian in the regional biostratigraphic scheme suggested for basins of northwestern Europe (Costa and Manum, 1988). Andreeva-Grigorovich included that species into name of the *Rhombodinium draco–Wetzeliella articulata* Zone of southern areas of the former USSR. *Systematophora placacantha* is the index species of late Lutetian subzone W9b of the Parisian basin. In the North Sea sections (Mudge and Bujak, 1994), the upper age limit of the species range corresponds to terminal Lutetian (middle and lower intervals of zones P12 and NP16, respectively). The *Systematophora placacantha* Zone characterizes the Lutetian Iset Formation penetrated by Borehole SG32 near the town of Kurgan in the northern continuation of the Turgai depression (Beniamovskii *et al.*, 1995). This formation is situated between the Irbit and Tavda formations like the green clay member with fish remains in the section of Borehole 9. In opinion of O.N. Vasil'eva, the zone in question spans the entire range of Lutetian Stage and zones W8 and W9 of the Parisian basin. It should be mentioned that the *Wetzeliella articulata–Systematophora placacantha* Zone is equivalent to intervals NP15–NP16 (lower part) and P11 (part)–P12 in zonations of nannoplankton and planktonic foraminifers, respectively. In Borehole 9, the green clay member is attributed to the *Pyxilla oligocaenica* Zone of diatoms, the range of later var. *tenuis* included. Simultaneously, it corresponds to the radiolarian *Lychnocanium separatum* Zone and possibly spans an interval of transition to the next *Heliodiscus quadratus* Zone. Kozlova, Strel'nikova, and Radionova suggest the late Ypresian–early Lutetian age of the *W. articulata–S. placacantha* Zone, while Glezer attributes it to the Lutetian Stage. Kozlova (1999) described the *L. separatum* radiolarian assemblage from the Kresty Formation of the Turan pate, where its post-Ypresian (likely late Lutetian) age is out of doubts. Like dinocysts, the *Acarinina rotundimarginata* assemblage of planktonic foraminifers from the lower interval of green clay member (depth level of 449 m in Borehole 9), which are characteristic of the Kuberle Horizon in the Crimea–Caucasus region, is of the Lutetian age as well. Planktonic foraminifers of the *Hantkenina alabamensis* assemblage (depth level of 445 m) elucidates age of middle strata of the green clay member (late Lutetian or, possibly, Lutetian–Bartonian transition period).

Based on planktonic assemblages, primarily on planktonic foraminifers and dinocysts, one can assume the growing influence of Tethyan waters on sedimentation environments in the basin. Taxa of different ages concentrated in the section interval of insignificant thickness suggest a general environmental reorganization during the Lutetian and transgression in the first half of that age, which is inferable from appearance of

planktonic foraminifers and diversification of dinocysts. The event was responsible for recurrent partial erosion of sediments, which were at the early diagenetic stage of consolidation. Signs of erosion are established at the depth levels of 451 m (the base of sandy layer below the clay member) and 446–444 m. Above the depth interval of 445 to 444 m, there are recorded signs of progressing regression and basin shoaling. They are consistent with compositional changes in diatom assemblages (data of Radionova) and with growing abundance of prasinophytes and acritarchs among the organic-walled phytoplankton (data of Zaporozhets). Simultaneously, dinocyst species became twice less diverse than before, and abundance of spores and pollen increased considerably, up to several times.

(c) *Terminal beds of the upper Lyulinvor Subformation* characterizing transition to overlying Tavda Formation are distinguished in the depth interval of 444–435 m. Dinocyst taxa characteristic of the initial middle Eocene (*Areosphaeridium diktyoplokum, Kisselovia ornata* f. *reticulata* and others) appear beginning from the depth of 445–444 m. Because of mass abundance of acritarchs and prasinophytes in the upper interval of the green clay member, it is difficult to define here the organic-walled phytoplankton zonation, and the interval is attributed to the *Paucilobimorpha–Micrhystridium* Beds. The beds are of ecological rather than biostratigraphic significance, because dinocyst assemblages from this level consist of euryhaline taxa of broad stratigraphic ranges. The same situation is typical of transitional beds between the Kresty and Kuma formations of the northern Caucasus, in which percentage of prasinophytes and acritarchs is as high as 70–80%. The event of shoaling in the West Siberian basin caused appearance of endemic forms among dinocysts (species *Kisselovia ornata* and *Thalassiphora elongata* unknown in western regions of Eurasia).

The basin shoaling that progressed toward commencement of the Tavda epoch influenced composition of the clay member that becomes enriched in silty and sandy components. Spores and pollen of the green clay member belong to the regional *Castanopsis pseudocingulum–Castanea crenataeformis* Zone established in upper Lutetian deposits of southern areas of the East European platform and Turan plate (Panova *et al.*, 1990). Thus, palynological data and microplankton suggest concordant ages of the green clay member.

The uncertainty of the Lutetian–Bartonian boundary and inconsistent zonations of nannoplankton and planktonic foraminifers within the transitional interval provoked controversial opinions with respect to that boundary position in the general scale. Berggren *et al.* (1995) believe that being 41.2 Ma old it should be at the base of chron C18r inside zones P12 (*Morozovella lehneri*) of planktonic foraminifers and NP16 of nannoplankton zonation. In the stratigraphic scale project recommended by J. Remane with colleagues at the International Geological Congress of 2000 in Brazil, the boundary is 2 m.y. older than in the scale of Berggren. Russian experts are of different opinions as well. In the scale of Interdepartmental Stratigraphic Committee (*Resolutions...*, 1989), it is placed at the base of the zones *Hantkenina alabamensis* (this is advocated by V.A. Krasheninnikov) and NP16. In the stratigraphic scheme authorized for the East European platform in 2000, the *Hantkenina alabamensis* Zone is placed into the Lutetian Stage, and the boundary is recommended to be defined based on nannoplankton.

In Borehole 9, the *Paucilobimorpha–Micrhystridium* Beds likely correspond to the terminal Lutetian. The inference is conventional however, as the Lutetian–Bartonian boundary position remains problematic until the competent decision of the International Subcommission of the Paleogene System. At present, it is reasonable to place this boundary at the base of the lower sandy member of the Tavda Formation (depth levels of 435 m in Borehole 9 and 456–455 m in Borehole 011BP). Being placed at the member base, the boundary appears to be well identifiable in logging diagrams and traceable over the plate.

The paleomagnetic zone N_6II established in the dinocyst zone *W. articulata–S. placacantha* of Borehole 9 can be correlated with the chron C20n only. In Borehole 011BP, a thin (3 m) subzone of reversed polarity at the top of zone N_1II may correspond to a part of chron C19r.

Tavda Formation

Lower Tavda Subformation

At the base of the lowermost sandy member of the Tavda Formation, there are recorded signs of erosion especially distinct in section of Borehole 011BP. The hiatus is insignificant however, as respective palynological spectra and dinocyst assemblages show characters inherited from underlying biostratigraphic units of the upper Lyulinvor Subformation.

The basal interval of the lower Tavda Subformation (depth range of 435–408 m in Borehole 9) corresponds to the *Areosphaeridium diktyoplokum(Rhombodinium draco* Zone of the early Bartonian. In the Bartonian Stage stratotype, the Elume Bay of White Island, southern England, the second index species occurs in association with nannoplankton of zone NP16 and with foraminifers of zone P13 at the level of 8 m above the stage base. Andreeva-Grigorovich and Zaporozhets (see in *Geological and Biotic...*, 1998) established the same situation in sections of southern Russia. In Borehole 9, the species range is controlled based on planktonic foraminifers found at the depth levels of 445 and 404–400 m. In the first case, species of the *Hantkenina alabamensis* assemblage, its index species included, define the lower limit (late Lutetian or Lutetian–Bartonian boundary interval), while *Acarinina rotundimarginata* present at the second level implies the Bartonian but not younger age of the regional zone in question, which is

conventionally attributed to the early Bartonian. Kul'kova who studied pollen and spores from the same interval attributed them to the *Quercus gracilis–Rhoipites granulatus–Castanea crenataeformis* palynozone. At the same time, Zaporozhets who found here *Castanopsis pseudocingulum* characteristic of the middle Eocene distinguished the studied interval as the *Quercus graciliformis–Castanopsis pseudocingulum* Beds.

Upper strata of the lower Tavda Subformation (depth range of 408–388 m) yield dinocysts of the *Rhombodinium porosum* Zone. In many regional dinocyst zonations of Boreal realm, this indicative zone is dated back to the late Bartonian. In the aforementioned stratotype of the Bartonian Stage, the zone has a concrete position being situated between the *Rhombodinium draco* and *Rhombodinium perforatum* zones. In England, but not in France, Germany, Belgium and Russia, the last zone is attributed to the late Eocene. As for correlation with biostratigraphic subdivisions of calcareous plankton, it corresponds in range to zones NP16 and P14 (incompletely as we think). In the dinocyst scale of Western Europe (Costa and Manum, 1988), it is at the level of zone D11. In the Russian and Turan plates, the *Rhombodinium porosum* assemblage is characteristic of the Saksaul'skii Formation of the late middle Eocene in the Aral and Turgai regions. In a sample from the depth level of 406.5 m (Borehole 9), Nikolaeva detected accumulation of ostracodes *Cyamocytheridea* aff. *corruguta* resembling *C. corrugata* characteristic of the Saksaul'skii Formation in the Manual'e section of the Irgyz area. Thus, the last group of fossils also points to the middle Eocene age of the lower Tavda Subformation. In the Voronezh anteclise, platform areas of Ukraine, and the northern Caucasus, the *Rhombodinium porosum* assemblage is confined to upper strata of the Kiev and Kuma formation, which are undoubtedly of the middle Eocene age. *Acarinina* and *Catapsydrax* forms found in the upper half of the lower Tavda Subformation suggest the same age, although the assemblage of planktonic foraminifers includes some species commonly occurring in transitional Bartonian–Priabonian strata. Like dinocysts, the foraminiferal assemblage originated in a relatively cooled basin of subtropical belt.

Two paleomagnetic zones of reversed (R_1tg) and normal (N_1tg) polarity distinguished in the lower Tavda Subformation of Borehole 9 can be correlated with chrons C18r and C18n. Characteristic of them are the *A. diktyoplokum–R. draco* and *R. porosum* dinocyst assemblages. In Borehole 011BP, paleomagnetic zones N_1tg, R_1tg, N_1tg yield the same assemblages and may correspond to chrons C19n, C18r, and C18n. In both sections, rocks with normal polarity prevail in the lower Tavda Subformation like in the correlative interval of the Berggren's scale.

Upper Tavda Subformation

Bearing dinocysts of the *Charlesdowniea clathrata angulosa* Zone, the upper Tavda Subformation, especially its upper half, can be directly correlated with the Priabonian of northern Europe and the Crimea–Caucasus region (zone D12 after Andreeva-Grigorovich in the last case). In sections of the Belaya, Kuban, and Kheu rivers (Zaporozhets, 1993, 1998, 2001), this dinocyst zone is correlated with zones of nannoplankton (NP18–NP20) and planktonic foraminifers (P15–P17). The same position of the zone is established in the Parisian basin, where it was originally established and dated back to the late Eocene. The dinocyst assemblage characteristic of the upper subformation facies deposited far seaward (depth interval of 295–277 m) is very diverse. It demonstrates a combination of "recurrent" taxa (*Areosphaeridium diktyoplokum, Enneadocysta arcuata, Phthanoperidinium eocenicum*, and others), which appeared in the middle Eocene and became extinct at the stage end, and transit taxa, which radiated widely in the Oligocene. Some cold-water species, e.g., *Phthanoperidinium amoenum*, probably appeared earlier in the West Siberian sea and then migrated in central and western areas of the Peri-Tethys during the early Oligocene cooling.

Ostracodes from middle and upper parts of the subformation (depth range of 343–291.8 m) also suggest the Priabonian age of host deposits. In addition to endemic *Cytherura placida*, the ostracod assemblage includes species known from the Chegan Formation of the Aral–Turgai region. Species *Clithrocytheridea innae* is widespread in the Khanabad and Sumsar sequences of Central Asia. The young transit form *Clytheridea* existed in the Oligocene as well. Older transit forms are *Paracypris contracta* (early–middle Eocene of the East European platform), *Cytheromorpha brabantica* (Lutetian–Bartonian of Belgium), and *Bythoceratina* ex gr. *impressa* (late Paleocene of Ukraine). Some of the listed taxa were found earlier in the Tavda Formation of West Siberia (Lipman and Nikolaeva, 1964).

A greater part of the Tavda Formation, the facies of freshened settings and upper transgressive member included, corresponds to the *Quercus gracilis–Quercus graciliformis* palynozone. Percentages of Fagaceae and other thermophilic taxa increase upward in the section, but basal beds of the overlying Isil'kul Formation are enriched in pollen of conifers (*Picea, Abies, Tsuga*), abundance of which is negligible in the Tavda formation. Spectra of amentiferous demonstrate an opposite trend: Betulaceae turn into dominants, Juglandaceae remain indifferent, and Fagaceae decrease in abundance. Palynological data represent the only biostratigraphic indication of a significant break in sedimentation and erosion of marine deposits.

Six paleomagnetic zones of normal and reversed polarity (R_2tg, N_2tg, R_3tg, N_3tg, R_4tg, N_4tg in Borehole 9 and R_2tg, N_3tg, R_3tg, N_4tg, R_4tg, N_5tg in Borehole 011BP) span the interval of the *Ch. clathrata angulosa* and *R. porosum* (partly) dinocyst zones, which are correlative with zones P15–P17 of planktonic foraminifers

(late Bartonian–Priabonian). According to this correlation, the distinguished paleomagnetic zones correspond to chrons C17r, C17n, C16r, C16n, C15r, and C15n. Rocks with normal polarity prevail in the upper Tavda Subformation.

PRINCIPAL BIOTIC AND ABIOTIC EVENTS OF THE TAVDA TIME

Paleontological data imply that sediments of the Tavda Formation penetrated by two boreholes recorded the same succession of biotic and abiotic events, which are important for the interregional correlation. Distribution trends of acritarchs and prasinophytes, diversity peaks of dinocysts, and corresponding periods of transgression maximums define the reference levels characterizing dynamics of sedimentation. In Borehole 9, a transition from shallow-water sediments to first transgressive facies is distinguished at the depth level of 412 m, where appearance of planktonic foraminifers and reduced abundance of *Paucilobimorpha* in phytoplankton are recorded. The early Tavda transgression maximum is established in boreholes 9 and 011BP (depth intervals of 408–392 and 430–410 m, respectively) based on appearance of planktonic foraminifers, diversification of dinocysts, and decreased abundance of prasinophytes and acritarchs. In both intervals, angiosperm pollen dominates in palynological spectra (70–80%), and Hamamelidaceae pollen is a considerable component. Transition from clayey to clayey–silty and silty sediments is recognizable at the depth levels of 388 and 401 m in boreholes 9 and 011Bp, respectively, and species diversity of dinocysts is reduced above both levels. Depth ranges of 388–367 and 401–368 m in two respective sections show transition from transgressive to regressive phase of the basin evolution. Abundance and diversity of dinocysts are decreasing here, and percentage of acritarchs and prasinophytes is insignificant. Indicative of the regressive phase is the mass abundance of aquatic fern *Hydropteris indutus* and parallel appearance of freshwater algae *Pediastrum* in some samples (depth intervals of 359–337 and 360–325 m, respectively). Pollen of Fagaceae is dominant in the *Quercus gracilis–Quercus graciliformis* Zone, the index taxa of which represent over 50% of the spectrum. Termination of the regressive phase corresponds to the biotic event recorded in depth ranges of 339–334 (Borehole 9) and 333.7–330.5, 325 m (Borehole 011BP). In shallow-water sediments of these levels, massulae and fronds of aquatic fern *Azolla vera* Krysht. occur in association with benthic foraminifers *Cribroelphidium rischtanicum*, ostracodes (*Cytheridea probata, Echinocytheris spongiosa, Paracypris compacta*), and bivalves (*Muscullus* cf. *samensis, Pitar circularis, Cardiopsis* sp.). According to information from S.V. Popov and I.A. Goncharova (Paleontological Institute RAS), the first and third taxa of bivalves are known only from the Chegan deposits of Ustyurt–Aral region, whereas the second one is characteristic of these and Oligocene deposits.

The next intervals in both sections characterize a gradual transition from regressive to transgressive cycle (330–313 m in Borehole 9 and 317.3–303.5 m in Borehole 011BP). Spores of aquatic ferns, prasinophytes, and acritarchs almost disappear from sediments at these levels. By the end of intervals, diversity of dinocysts increases up to 30 species. Cooling of that time is evident from a reduced abundance of angiosperm pollen, especially of thermophilic taxa, and from a higher percentage of Juglandaceae and Betulaceae (5%) in samples 330, 326, and 313. The Tavda transgression maximum corresponds to depth intervals of 394–277 and 298.7–265 m in boreholes 9 and 011BP, respectively, where the diverse dinocyst assemblages (50 to 55 species) include recurrent forms of the lower Tavda Subformation. Representatives of the cold-water affinity from genera *Rhombodinium* and *Phthanoperidinium* become frequent here. Fagaceae dominate again in spore–pollen spectrà.

Thus, the established trend of biotic and abiotic events of the Tavda time appears to be useful for interregional correlation (Table 1). The trend depicts the mass appearance and subsequent quick reduction of acritarchs and prasinophytes, fluctuations in abundance and diversity of dinocysts, and episodic simultaneous appearance of other groups of marine biota (transgressive pulses?). The most remarkable is transition from regressive to transgressive phase of the mid-Tavda time, when massulae and fronds of *Azolla vera*, benthic foraminifers of the *Cribroelphidium rischtanicum* assemblage, ostracodes (*Cytheridea probata, Echinocytheris spongiosa, Paracypris compacta*), and bivalves (*Musculus* cf. *samensis, Pitar circularis, Cardiopsis* sp.) were buried in calcareous sediments of a narrow interval (3 to 5 m thick).

The early and late Tavda transgressive episodes are contrastingly recorded in two sections. Being hardly detectable in Borehole 9, transition to the reduced half-cycle of the late Tavda time is unrecognizable in Borehole 011BP.

Sedimentation rate of the Tavda time (27 mm per 1000 years in average) was much greater than in the Lyulinvor period. It was initially relatively low (10 mm per 1000 years) and increased up to 37–39 mm per 1000 years in the late Tavda time. The plate downwarping did not compensate the intense influx of terrigenous material, and sea basin experienced a considerable shoaling in the mid-Tavda time. During the transgression peaks, the basin could be 100 and more meters deep in accumulation settings of fine clayey sediments only.

Gray kaolinic silty clays of the Isil'kul Formation (age equivalent of the Atlym Formation in the Central lithologic-facies zone) rest on uneven eroded surface of underlying green clays of the Tavda Formation. Small, flattened pebbles of green clay are widespread in basal

Table 1. Main stages of the Tavda basin evolution

Main stages of the basin evolution	Principal indications of biotic and abiotic events	Depth interval (m); Borehole 9	Depth interval (m); Borehole 011BP	Additional information
Termination of regressive stage that antedated the early Tavda transgression	Gradual replacement of sandy–silty deposits by clay; reducing abundance of *Paucilobimorpha* and other acritarchs and prasinophytes	435–412	455–430 (?)	Offshore distance coefficient 10 to 15%
Transgressive phase of the early Tavda time	Accumulation of clayey sediments; appearance of planktonic foraminifers and ostracodes, growing abundance and diversity of dinocysts	408–392	430–410	Offshore distance coefficient up to 25%; warming
Transition from transgressive to regressive phase	Replacement of clayey sediments by silty deposits; decreasing content and diversity of dinocysts and growing abundance of acritarchs and prasinophytes	388–367	401–367	Drop of offshore distance coefficient down to 15–18%
Regressive phase of the mid-Tavda time	Prevalence of silty sediments; further decrease of dinocyst diversity and mass abundance of *Hydropteris, Pediastrum*, and thin-walled dinocysts *Hysrtichokolpoma* and others	359–337	360–325	Further decrease of offshore distance coefficient down to 10–12%
Biotic episode of the incipient transition from regression to transgression	Higher carbonate content in sediments; mass abundance of *Azolla* and *Hydropteris*, accumulations of foraminifers, ostracodes, and mollusks	339–334	333.7–330.5, 325	Resumed growth of offshore distance coefficient
Transition from regressive to transgressive phase	Alternating clay, sand, and silt interlayers, siderite lenses; gradual diversification of dinocysts and higher content of coniferous pollen with parallel reduction of thermophilic angiosperms in palynological spectra	330–313	317.3–303.5	Slow increase of offshore distance coefficient; cooling
Transgressive phase of the late Tavda time	Prevailing clay beds intercalated with sand, silt, and siderite interlayers and lenses; abundance growth and diversification of dinocysts (up to 60 species); palynospectra with thermophilic taxa and dominant gymnosperms	304–277	298.7–265	Offshore distance coefficient up to 25–30%

Isil'kul beds. Palynological assemblages below and above the erosion surface are very different, implying a considerable break in sedimentation. Dominant Fagaceae and Taxodiaceae occurring in association with thermophilic *Engelhardtia, Corylopsis, Hamamelis, Ilex*, and *Palmae* are taxa characteristic of pollen spectrum from the depth level of 277 m, whereas Pinaceae and Betulaceae dominate in pollen at the depth of 274 m.

Paleomagnetic data also suggest a considerable hiatus between the Tavda and Isil'kul formations. The equivalent of chron C13r of the Berggren's scale is missed from the studied sections of Paleogene deposits. Magnetostratigraphic records near the Eocene–Oligocene boundary, which is established in most complete sections of Mediterranean and Crimea–Caucasus regions based on calcareous nannoplankton, are complicated, consisting of normal and reversed polarity chrons of variable durations and succession. In such a situation, it is difficult to correlate the general and regional magnetostratigraphic scales. In the general scale, the reversed polarity chrons 13r and 12r are separated by interval of normal polarity (13n) between 34.6 and 30.9 Ma, which spans about one third of the Priabonian and the greater part of the Rupelian. In boreholes 011BP and 9, the upper transgressive beds of the Tavda Formation correspond, in opinion of Gnibidenko, to the respective normal-polarity chron in the regional magnetostratigraphic scale of West Siberia. In the general scale, they may span the interval of paleomagnetic chrons C16 and C15.

ONCE MORE ABOUT THE TAVDA FORMATION AGE

Despite the diverse viewpoints on the Tavda Formation age, there is a number of undisputable observation results.

First, nobody argues now against the two-member structure of the Tavda Formation composed of two transgressive–regressive sedimentary cycles in most complete sections of central West Siberia and its peripheral zone. Transgressive facies of the cycles correspond to clayey members, the principal ones in structure of the lower and upper Tavda subformations.

Second, the lower Tavda Subformation is naturally limited by sand or sandstone members detectable in logs as regional logging markers "T" and "B." In those cases, when sandy character of sediments in the Tavda Formation middle is not obvious, like for instance in boreholes 9 and 011BP, paleontologists detect the corresponding level based on appearance of brackish-water mollusks, abundant acritarchs and prasinophytes, aquatic fern spores. It is also important that in latitudinal profiles, which cross Paleogene deposits along and on both sides of the Omsk–Novosibirsk railway, continental coaliferous facies replace marine sediments beginning exactly from the logging marker "B."

Third, all fossil groups used to define the Tavda Formation age (ostracodes, Elasmobranchii, benthic foraminifers, mollusks, dinocysts) include many taxa, which are characteristic of the following stratigraphic subdivisions in other regions: the Chegan Formation of Ustyurt, Aral and Turgai regions; Adaevka Formation of Mangyshlak; Rishtan, Isfara, Khanabad, and Sumsar horizons of marine Paleogene deposits in the Fergana and Tajik depressions of Central Asia. Experts on foraminifers (Subbotina, 1937; Lipman, 1955; Ushakova, 1957) and other paleontologists (Boitsova *et al.*, 1968; Khokhlova and Nikolaeva, 1964; Panova, 1967) already paid the necessary attention to this fact. Kisel'man (1978) who compared benthic foraminifers from Central Asia and Tavda Formation figured out that 9 of 16 elphidiid, nonionid, miliolid, and anomalinid species were discovered first in the Tavda Horizon, while the other seven taxa were defined in Paleogene deposits of Central Asia. "*Cribroelphidium rischtanicum* appears in the Rishtan strata and disappears at the Sumsar level; *Discorbis ferganensis* and *Nonion laevis* are detected in the Rishtan and Khanabad beds, being unknown in Fergana above the latter. The most abundant species of the *C. rischtanicum–Brotzenella munda* (N. Bykova) assemblage is simultaneously widespread in Khanabad and Sumsar strata of Fergana (Bykova, 1939). *Nonion graniferus* Terquem is also confined to the Khanabad level ... ... Since *C. rischtanicum* assemblage includes some polymorphinids, abundant miliolids, and *B. munda*, it is close to assemblage of the Khanabad regional stage" (Kisel'man, 1978; pp. 45, 46). Admitting then that *C. differensapertio* subassemblage, the upper one among three subassemblages distinguished by M.V. Ushakova in the *C. rischtanicum* Beds, may correspond as well to a higher Sumsar level of Central Asian sections, Kisel'man concluded nevertheless that this level "... can be correlated with the lower–middle interval of the Sumsar Beds in Central Asia." It is also important that Kisel'man (like Ushakova before) defined the appearance of *C. rischtanicum* assemblage at the level of 25–30 m (Tyumen borehole 1-P) and 120 m (Tar profile) above the base of Tavda Horizon. In our opinion, these data suggest, though indirectly, that lower beds of the Tavda Formation are older than the late Eocene.

Taking into account recent age determinations of the Sumsar Beds, we tried to define the Tavda Formation age based on subdivisions of the regional Paleogene scale of Central Asia. This was reasonable in view of general paleogeographic situation in Eurasia during the late Eocene, when the West Siberian sea, Tajik and Fergana basins represented eastern gulfs of the Peri-Tethys, in which eustatic movements were comparable.

It is important to compare the West Siberian and Central Asian bio- and lithostratigraphic units, because the Sumsar Beds of the Tajik and Fergana depressions bear nannoplankton (Muzylev and Salibaev, 1988; Muzylev *et al.*, 1996). In the nannoplankton assemblage from the upper Kushan (Khanabad) and Sumsar

deposits of the Tajik depression, Muzylev identified single specimens of *Isthmolithus recurvus*, the index species of zone NP19/20. In the Sumsar Beds, nannoplankton of this zone is more abundant and diverse. In the Fergana depression, the other index species *Discoaster barbadiensis* was found in upper marine deposits of the Sumsar sequence stratotype near the town of Isfara. The nannoplankton assemblage characteristic of zone NP19/20 is also known from the Shor-su section of the Fergana depression. In addition to above index species, the assemblage includes *Rhabdosphaera spinula, Rh. crebra*, and *Braarudosphaera bigelowi*. Muzylev termed it the *Istmolithus recurvus–Sphenolithus pseudoradians* assemblage of undivided zones NP19/NP20 of the terminal late Eocene.

Muzylev *et al.* (1996) also reported that in the Fergana depression the Sumsar Beds transgressively overlap the eroded surface of older deposits, the Turkestan strata included, as it was first noted by Makarova and Mironova (1964) and then by Salibaev (1982). Thus, the upper Eocene of Central Asian depressions is represented by two (but not one) transgressive–regressive cycles, one of the late Rishtan–Khanabad and the other one of the Sumsar time. Transition from the lower Rishtan sands to upper Rishtan slightly calcareous clays marks the commencement of the first cycle. It was a conventional decision to regard the middle–upper Eocene boundary as coinciding with boundary between the cycles, because nannoplankton of upper Rishtan deposits is lacking index species. In opinion of Muzylev, appearance of *Crubrocentrum reticulatum* "implies that host deposits are not older than the late middle Eocene; the event does not specify stratigraphic position of the Rishtan Horizon being consistent with available viewpoints" (Muzylev *et al.*, 1996, p. 42). The regression peak of the first cycle corresponds to the pre-Sumsar hiatus. The second (Sumsar) cycle is reduced, represented by transgressive half-cycle only (calcareous clays with abundant nannoplankton), and its regressive part is consumed by the pre-Massaget hiatus. Muzylev and his colleagues correlated the mentioned half-cycle with the late Priabonian (Lattorfian) eustatic transgression. The same situation is established in the Aral–Turgai region of the Turan plate, where the regressive half-cycle of the Chegan sedimentary cycle is also reduced, and Eocene assemblages of foraminifers and dinocysts are close in composition to Central Asian and West Siberian counterparts.

In this connection, we should remind that second (late Tavda) sedimentary cycle is actually represented in studied sections of West Siberia by the lower transgressive part, while the regressive half-cycle corresponds either to the Kurgan Beds (if they are deposited during the quick shoaling of the Tavda basin) or to the pre-Atlym hiatus. Anyway, the uniform eustatic trends and great similarity between the Tavda and Rishtan–Sumsar biotas suggest definitely, as we think, the late Eocene age for the upper Tavda Subformation. In other words, the interregional correlation supports conclusions based on studied fossil groups, which are less dependent on the facies control (Elasmobranchii, dinocysts, spores, pollen, fructifications of terrestrial plants, and some ostracodes).

REGIONAL CONSIDERATIONS AND ANALYSIS OF MARINE PALEOGENE SECTIONS IN THE BARABINSK LITHOLOGIC-FACIES ZONE (DRILLING RESULTS OF 1950–1970)

Now, after discussion of age problems, we should understand to what extent biostratigraphy of studied sections is applicable to other areas of the Barabinsk and surrounding Ishim, Central, Narym, Ob, Kulunda and Pavlodar lithologic-facies zones, where deposits of the Lyulinvor and Tavda formations are widespread (*The Unified...*, 2001). First, we should consider data on nearby boreholes, which penetrated throughout the Paleogene sequence in these zones.

In the Barabinsk zone of West Siberia, profiles and network of exploration and Craelius drilling with associated logging was in progress since the 1950s. A series of boreholes drilled in the Tatarskoe, Ipatovo–Svetlinskaya, Fedorovskaya, Burlinskii and other exploration areas elucidated the correlation potential of lithostratigraphic boundaries detectable in logs. Dozens of boreholes, which penetrated down to the top of Mesozoic sequence and deeper (Omsk, Barabinsk, and other sites), have been sampled and studied in detail (paleontological analysis included) and can be regarded as the Paleogene reference sections. Analyzing drilling results published in the 1950s and 1960s, we should bear in mind some necessary corrections. For instance, we place the bed of glauconite sand (the logging marker "T" between the Lyulinvor and Tavda formations) at the base of the Tavda Formation, as it was done by Martynov (1967, 1968, 1973), Zal'tsman (1968) and Shatskii (1978), although earlier, in 1955–1965, it was regarded as terminal bed of the Lyulinvor Formation. In general, ages of lithostratigraphic units were based at that time on a few groups of fossils, and elaboration of zonal scales for most informative groups just started. T.F. Vozzhennikova did only first steps toward a comprehensive investigation of organic-walled phytoplankton. Stratigraphic significance of planktonic foraminifers and Elasmobranchii was underestimated, and main attention was paid to benthic foraminifers, siliceous microplankton, ostracodes, and palynomorphs of terrestrial plants. Facing the undeveloped state of the general stratigraphic scale for Paleogene, where even subseries ranges were inadequately substantiated, paleontologists referred in their works to subdivisions of the regional scale authorized for the Crimea–Caucasus region. In particular, they correlated the greater part of Lyulinvor Formation with the Bodrakian regional stage and thus attributed it to the upper Eocene. As is known now, this stage corresponds to the middle Eocene, i.e., to the Bartonian and partly Lutetian stages (the Kuberle Horizon at least). Accordingly, we had to correct the

Table 2. Principal characteristics of the marine Paleogene sections recovered by reference boreholes in the Barabinsk lithologic-facies zone of the West Siberian plate

Position of markers and thickness of lithostratigraphic units	Boreholes			
	2	3	4	5
Depth to the Tavda Formation top, m	274	254	317	299
Depth to the Tavda Formation base and thickness of basal sand bed (logging marker "T")	435 426–435 9	443 436–443 9	475 465–475 10	434 432–434 2
Thickness of the Tavda Formation, m	162	189	158	165
Depth to the horizon "B" between lower and upper Tavda subformations (after Gurari, Martynov, Shatskii and others) and its thickness, m	?	354.3–358 3.7	400–404 4	360 ?
Depth to the horizon "B" and its thickness (m) in the lower part of the Tavda Formation (after Zal'tsman, 1968)	? 388.3–388.5 0.5	406.2–410.4 4.2	Not detected	Not detected
Depth to the base of upper Lyulinvor Subformation; presence (+) of basal horizon of glauconite sand and its thickness, m	+ 491–492.5 1.5	+ 497–499.75 2.75	547 ?	480 ?
Thickness of the upper Lyulinvor Subformation	57	54.5	? 80	46
Depth range and thickness (m) of green clay member of the upper Lyulinvor Subformation	450.9–449.9 1.0	461–466 5.0	481–496 15	?
Depth range and thickness (m) of sand interlayers inside the green clay member	447.5–448 0.5	Not detected	Not detected	Not detected
Presence of fish remains (+) and diatoms of the *Pyxilla oligocaenica* var. *tenuis* Zone (x) in the green clay member	+ x	+ x	? ?	? ?
Depth range and thickness (m) of glauconite-sand bed at the base of lower Lyulinvor Subformation	519–521 2	575–575.15 2.15	598 ?	512–522 1
Recovery range and thickness (m) of lower Lyulinvor Subformation	499.5–521 28.5	497.5–575.15 79.65	547–597 50	480–522 42
Total thickness of Lyulinvor Formation, m	86	132	130	88
Depth range and thickness (m) of the Talitsa Formation	Unknown	575–591 16	597–607 10	Unknown
Depth to the top of Mesozoic Gan'kino Formation, m	521	591	607	522

former understanding of litho- and biostratigraphic subdivisions of the Paleocene, lower and, especially, upper Eocene.

The issues we paid attention to during comparison and consideration of our results in the regional aspect were as follows:

(1) Relationships between the Cenozoic base and underlying Mesozoic deposits, the Gan'kino Formation included;

(2) Presence or absence of the Talitsa Formation in the southeast of West Siberia;

(3) Erosion marks in sediments underlying the horizon of glauconite sandstone with phosphorites, the basal one in the lower Lyulinvor Subformation;

(4) The same characters near the base of the upper Lyulinvor Subformation;

(5) Interlayers of sandy–silty glauconite-bearing sediments at the base of green clay member of the "Chegan" type (Nyurol'ka Formation s. str. after Shatskii), which are detectable in some logs (e.g., Borehole 9) and separate that member from the main upper Lyulinvor sequence composed of gray opoka-like and diatomaceous clays; their presence is significant therefore in view of hiatus at the Ypresian–Lutetian boundary level established in Borehole 9;

(6) Logging markers, which may correspond to similar interlayers inside the green clay member, and their correlation with breaks in sedimentation like the one found in section of Borehole 9;

Boreholes						
6	10	11	9	12	7	8
292	268	308	305	254	273	35
394	360	468	485	352	367	104.5
388–394 / 6	352–360 / 8	464–468 / 4	472–485 / 13	344–352 / 8	360–367 / 7	101.5–104 / 2.5
102	92	160	180	98	94	64.5
322.5–328.1 / 5.6	?	307–309 / 2	388–390 / 2	274–285 / 11	307–309 / 2	67.5–68.5 / 1
Not detected	326–331 / 5	423–427 / 4	?	313–315 / 2	333–336 / 3	?
	+				+	+
?	378–380 / 2	492 / ?	?	?	389–390 / 1	147–148.3 / 1.3
–						
? 43	20	24	?	35	23	43.8
409.5–414.8 / 5.3	370 / ?	478–480 / 2	503–506 / 3	362.5–363 / 0.5	?	111–112.5 / 1.5
Not detected	Not detected	Not detected	Not detected	Not detected	Not detected	Not detected
+	+	?	?	+	?	+
?	?	?	?	x	x	x
464 / ?	378–380 / 2	544–542 / 2	575–580 / 5.0	400–405 / 5	417–419 / 2	160.5–161 / 0.5
437–464 / 27	380–411 / 31	492–544 / 52	Not detected	387–400 / 13	30–419 / 29	148.3–161 / 12.7
70	51	76	95	53	52	56.5
464–468 / 4	Unknown	544–551 / 7	Unknown	Unknown	Unknown	Unknown
468	411	551	580	405	419	161a

(7) Data on coarser grain size in topmost strata of the upper Lyulinvor Subformation, as they indirectly suggest the basin shoaling, and on beds with diatoms of the *Pyxilla gracilis* var. *oligocaenica* Zone, radiolarians of the *Lychnocanium separatum* assemblage, and dinocysts replacing taxa of the *Charlesdowniea clathrata* s.l. Zone;

(8) Fish remains and planktonic foraminifers, if they present in the green clay member and may be regarded therefore as additional correlation criterion;

(9) The lateral extent of the first sandy horizon at the Tavda Formation base (logging marker "T") and indications of hiatus below this level;

(10) The lateral extent of sandy horizon (logging marker "B") inside the Tavda Formation, which was discriminated first by F.G. Gurari and then by Martynov and Shatskii, being regarded as the boundary level between two subformations;

(11) Data clarifying age of upper Tavda horizons and indications suggesting the transgressive character of the uppermost clay member.

We analyzed the necessary information compiling data on well-documented reference boreholes, those of the Barabinsk and Omsk sites included, and summary of collected materials is presented in Table 2. Geographic localities of boreholes we referred to are shown in Fig. 1.

Like geologists who mapped the region, we divide the Lyulinvor Formation of the Barabinsk zone not in three, but in two subformations, and this approach is acknowledged in the Unified regional stratigraphic scheme of 2001. It was easy therefore to compare

lithostratigraphic units of studied boreholes 9 and 011BP with correlative subdivisions discriminated by Martynov (1967, 1968, 1973) and Zal'tsman in the south of the West Siberian plate, as one can see from information on above issues (Table 2), especially on those numbered 1–4 and 9.

Let us consider first the problem of Talitsa Formation in the Barabinsk and adjacent zones. As is clear from Table 2, thin sequences comparable with that formation can be discriminated in sections of the Omsk, Barabinsk, and Tatarskoe reference boreholes in the northernmost area of the zone. Zal'tsman (1968, p. 14) tentatively discriminated the Talitsa Formation at the base of the Paleogene in Borehole 32, the village of Mikhailovka of the northern Kulunda area. His decision was misleading, as we think, because the corresponding 7-m-thick member (core recovery 1.2 m only) is composed of alternating shales and siltstones, which are lacking marine fauna and yield plant remains only. Similar continental deposits are widespread along periphery of the West Siberian depression, e.g., in the Zaisan basin, where they are of the same composition and yield Danian flora of the Tsagayan type. Palynological spectrum of Danian beds in that basin (*Practical Palynology*, 1990) is identical to spectra characteristic of continental terrigenous deposits penetrated by boreholes in northern and eastern flanks of the Kazakh Upland (Kopytova *et al.*, 1960; Antypko, 1962; and others). In dozens of other boreholes drilled in the Barabinsk (central and southern areas) and Kulunda zones, the Lyulinvor Formation rests directly on the eroded surface of Cretaceous deposits. It seems that the Talitsa sea occupied only a narrow boundary territory between the Barabinsk and Central zones. In southern areas of the Central zone, e.g., in the Tar reference section, the Talitsa Formation of little thickness is composed of typical marine sediments.

The limited size of the Talitsa sea predetermined transgressive onlap relations between the basal coarse-clastic horizon of glauconite-phosphorite sandstone of the lower Lyulinvor Subformation and underlying Cretaceous deposits. As one can see from Table 2, this horizon has been penetrated by all reference boreholes without exception. Even if the proper interval has not been sampled, the horizon is clearly detectable in logs. It has been penetrated everywhere in the junction area of the Barabinsk and Kulunda zones at the depths of 523 and 500 m in boreholes 3 and 23, Trofimovka area, at the depth of 510 m in Borehole 123, Berezovka area (50 km southward of Borehole 9), and in Borehole 111, Zhelezinka area, the Irtysh River right side. Boreholes of southern Pavlodar–Kulunda profile (Uspenka 35, Kulunda 14, Efremovka 185 and 48) and those situated in central and eastern Kulunda areas (Alekseevka, Acyamovo, Slavgorod and other sites) also proved presence of the basal sandy horizon under consideration. Martynov (1973) and Zal'tsman (1968) showed that the Lyulinvor Formation base overlies in places Maastrichtian or upper Campanian strata of the Gan'kino Formation and even lower Campanian horizons (Slavgorod Formation), as in central Kulunda areas, for instance. Characteristic of the Lyulinvor Formation itself is persistent alternation of dark-colored opokas, siltstones and sandstones of quartz–glauconite type, which is established everywhere.

In the Trans-Urals zone of West Siberia, the Thanetian age of the Serov Formation (age analogue of the lower Lyulinvor Subformation) is convincingly substantiated based on siliceous microplankton and dinocysts, whereas palynofloras played in contrast the prime role in correlation and age determination of host deposits in southeastern lithologic-facies zones. The studied dinocysts from Borehole 9 proved for the first time the Thanetian age of the lower Lyulinvor Subformation and showed that our colleagues rightfully regarded this subdivision as concurrent in age to the Serov Formation.

In distinction from lower subformation, discrimination of middle and upper subdivisions of the Lyulinvor Formation was always problematic. In the Barabinsk zone, where formation was divided in two subformations, its terminal green clay member was included into the upper Lyulinvor Subformation together with underlying member of gray opoka-like clay and mudstones, as it is done, for instance, in documentation of core samples from boreholes 9 and 011BP. In northerly areas, closer to the Vasyugan River basin, the upper subformation included the terminal member only, while the underlying gray beds (diatomaceous opoka-like clay, mudstones, and opokas) were attributed to the middle subformation (Kriventsov, 1984). In addition to formerly different nomenclature of the same lithostratigraphic units, we should also mention that there are coarse-grained horizons and erosion marks below them at basal levels of the upper Lyulinvor Subformation (as it is ranked in the Barabinsk zone) and its terminal member. Beds of this kind are often detectable also in borehole logs, as one can see in Table 2. Similar data are known as well in southern and northern areas of West Siberia (Martynov, 1967, 1968, 1973; Zal'tsman, 1968; Antypko, 1962; Dergachev and Lysenko, 1984). Because the green clay member penetrated by Borehole 9 yielded planktonic foraminifers, it is appropriate to mention here data elucidating the character of contact between this unit and underlying gray diatomaceous and opoka-like clays in the Vasyugan River basin, the stratotype area of the Nyurol'ka Formation distinguished by Shatskii. After geological survey in this area (Nyurol'ka and Chizhapka rivers, Ob River left tributaries upstream of the Vasyugan River mouth), Kriventsov (1984, p. 89) wrote: "The upper Lyulinvor sequence (Nyurol'ka Formation after Shatskii) is composed of clay that is greenish gray when naturally wet to greenish yellow in dry state, compact or platy, with finest laminae of silt, mica, or microscopic plant detritus. At the contact with underlying deposits, there are erosion marks, laminae of sand, separate quartz pebbles and clay pellets." He observed the same peculiarities in

sections of boreholes 5k and 45p drilled in the Chuzik River basin and was first to note that the marine Paleogene sequence is of maximum thickness (over 300 m) precisely in the Nyurol'ka depression. Considering arguments in favor of hiatus below the base of green clay member, we should bear in mind as well the two points of interest. First, development of transgression usually results in erosion of basal transgressive facies and contact with underlying sediments may correspond to a cryptic hiatus, and second, sandy beds of insignificant thickness may be washed out by drilling and removed from borehole by flushing. Nevertheless, many logs show that originally they were in the sequence. Thus, the discrete age of green clay member and indications of erosion at its base may be taken for arguments in favor of rehabilitation of the Nyurol'ka Formation.

Situated at the Tavda Formation base, glauconite sand and sandstone beds with erosion marks below them (logging marker "T") are widespread in southern areas of West Siberia (see Table 2), especially in the marginal plate zone adjacent to fold structures of Kazakhstan, Altai, and the Salair Range. Borehole 3 near the village of Trofimovka (Kulunda zone) penetrated bed of this kind, which is 10 m thick, and similar beds were detected at the Kulunda, Slavgorod, and other sites (Zal'tsman, 1968). Numerous boreholes drilled in Central and Narym zones outlined distribution areas of the Tavda basal sands in the Ob–Irtysh interfluve and Vasyugan River basin (Shatskii *et al.*, 1973; Kriventsov, 1973, 1984). Deposition of these coarse-grained sediments postdated shoaling of the Lyulinvor basin, which is evident from changes in diatom and dinoflagellate assemblages and from mass abundance of acritarchs and prasinophytes (*Paucilobimorpha, Micrhystridium*) at the corresponding level. The basin remained shallow until the commencement of the early Tavda transgression. The Lyulinvor–Tavda transition is not uniform over West Siberia. In the Central zone, it is gradual, demonstrating transition from green clay beds to the so-called Taichi member (Martynov, 1973) of alternating sandstones, siltstones and shales, and then again to clay beds prevailing in the Tavda sequence. In places, break in sedimentation was insignificant, and Lutetian regressive beds (green clays) of the Lyulinvor sedimentary cycle are preserved in the sequence to a considerable extent (Borehole 9). In other cases nevertheless, these beds are eroded completely or almost completely (Borehole 011BP), and the basal sandstone bed of the Tavda Formation directly overlies the gray member of diatomaceous clays. In marginal areas of the West Siberian basin, the pre-Tavda hiatus can be larger, corresponding in range to two microplankton zones (Iakovleva, 1999).

OCEANIC INFLUENCE ON SEDIMENTATION IN THE WEST SIBERIAN INNER SEA

Complicated relationships between the Lyulinvor and Tavda formations originated in response to changeable hydrological and sedimentation regimes in the West Siberian basin. The Ypresian period of accumulation of gray diatomaceous and opoka-like clays of the upper Lyulinvor Subformation corresponded to the time, when sedimentation was under great influence of arctic waters and the inner sea was connected with the World Ocean. This is evident from organogenic opal present in sediments depleted in the heavy mineral fraction.

During the early Lutetian, when accumulation of the green clay member commenced, biota and sedimentation in the basin experienced influence of both the northern and southern oceanic waters, as the basin was of maximum size at that time of the high sea-level stand in the World Ocean. Later on, when effect of arctic waters became weaker, influence of the Tethys increased, and planktonic foraminifers appeared in the basin. In addition, the growing impact of southern connections conditioned participation of carbonate components in sedimentation, decrease and subsequent disappearance of organogenic opal from sediments, which became enriched in glauconite, chlorite, and epidote responsible for their green coloration. The arctic influence completely ceased in the Bartonian time, especially in southern areas of the plate, as it is recorded across the Lyulinvor–Tavda boundary. Tethyan waters influenced sedimentation of this time not so intensely as well, because the late Eocene was a period of general sea-level drop in the World Ocean. Two transgressions of the Tavda period advanced however from the south, and both caused increase of sedimentation rates and higher contents of chlorite and epidote in sediments. These indications suggest a noticeable shoaling of the basin rather than crustal subsidence, and compensated sedimentation reduced the basin depth from hundreds to a few dozens of meters in average. The much lower "offshore distance coefficients" calculated for respective palynological spectra are indirect arguments in favor of this process.

The logging marker "B" and corresponding layer of glauconite sand between the lower and upper Tavda subformations can be diachronous in different lithologic-facies zones, as the marker is at a lower stratigraphic position in southern sections. At the same time, the Tavda Formation is of a distinct three-member structure in southeastern areas of the West Siberian plate, especially in the Kulunda zone, where its lower and upper transgressive clay beds are separated by the middle member of siltstones deposited in shallow settings, as established in sections of boreholes 9 and 011BP. In the eastward direction, these shallow-water sediments grade into continental coaliferous deposits. Farther to the east, rocks of the Tavda formation are completely replaced by continental sediments of the Yurkovo Formation. In their interpretation of the above facies trend, Gurari and Shatskii suggested that after the middle member continental deposits of eastern areas replace clays of the lower Tavda Subformation and all sediments constitute here a single continental

sequence. In their opinion, closer to the west the formation is of the two-members structure with a thin wedge of upper transgressive clays, which pinch out in easterly areas. The contact between clays of the upper Tavda Subformation and overlying gray sandy–silty deposits of the Isil'kul (Atlym) Formation is of a distinct transgressive type, indicative of erosion below it, and the subformation is represented in studied sections by its transgressive half-cycle only.

CLIMATE

Paleocene and Eocene climatic fluctuations in southern areas of West Siberia depended on global and regional factors. The latter, i.e., the character of the inner epicontinental sea connections with the Arctic and Tethyan oceanic basins controlled the paleotemperature and humidity fluctuations, the annual precipitation trends included. Indicators of relatively high temperature of near-surface waters in the sea, especially during the early Eocene, are the high diversity of microplankton, foraminifers of genera *Acarinina* and *Hantkenina*, which are found in sediments and characteristic of the Earth subtropical areas, some radiolarians also typical of these areas (e.g., species of the *B. clinata–B. longa* Zone), and chorate cysts prevailing in phytoplankton. The latter dwell in seas, where temperature of the upper photic zone is up to 20°C. Terrestrial vegetation is even more sensitive to climatic changes. As is known for a long time, palyno- and macrofloras from marine or alternating marine and continental Paleogene strata of the study region include several groups of dominant taxa, which successively replace one another in the section and have been used to denote the respective zonal assemblages. For instance, Fagaceae pollen of the *Castanea crenataeformis–Castanopsis pseudocingulum* palynozone is very abundant in the late Ypresian–Lutetian strata. In Bartonian deposits, Fagaceae are of another taxonomic composition, and pollen of above palynozone occurs in association with smaller pollen of oaks of the *Quercus gracilis–Quercus graciliformis* group. The Priabonian sediments yield only pollen of the last group.

The same succession of generic taxa is characteristic of macrofloras from the Pavlodar area. Here, the Ypresian–early Lutetian Karasor flora with dominant broad-leaved *Castanopsis* and Lauraceae forms gives way upward to Bartonian Takyrsor, Zhamantuz and other xeromorphic subtropical floras, in which small-leaved sclerophyllous oaks are dominant taxa (Makulbekov, 1972, 1997; Akhmet'ev *et al.*, in press).

In the present-day vegetation cover, *Castanea* and especially *Castanopsis* forms (over 100 species in total) are widespread in tropical and subtropical areas, mainly in Southeast Asia with its monsoon climate and precipitation maximum in summer time (Burma, Vietnam, southern and southeastern China, southern Japan). Recent counterparts of sclerophyllous oaks are also characteristic of seasonal climatic areas, though of the Mediterranean type in this case, with precipitation peaks during the winter season. Besides the Mediterranean, oaks of this kind are widespread in the southwestern United States and adjacent areas of Mexico. The monsoon climate of the early Eocene developed in the study region because of through connections between the Tethys and Arctic ocean via the Turgai strait and West Siberian sea. Hindering of connections in the terminal Lutetian and earliest Bartonian gave rise to a wide development of xeromorphic sclerophyllous oaks in the vegetation cover. By the terminal Bartonian and in Priabonian, when the West Siberian sea was completely isolated from the Arctic Ocean, the Mediterranean climate replaced the former one with summer monsoons. At that time, the West Siberian basin looked like a giant gulf extending far northward from the Tethys marginal zone.

An insignificant mid-Priabonian cooling determined a higher percentage of coniferous, Betulaceae, and Juglandaceae pollen in palynological spectra. The reduced content of gymnosperm pollen in the upper 30-m-thick sequence of the Tavda Formation in Borehole 9 and concurrent abundance growth of evergreen deciduous angiosperm taxa of thermophilic affinity imply that subsequent annual temperature increase of 2–3°C was connected with transgression of the late Tavda time, when climate became more humid. The respective period of subtropical climate lasted until the end of Eocene.

CONCLUSION

(1) Two reference boreholes drilled in the Barabinsk lithologic-facies zone penetrated through the marine Paleogene sequence of the West Siberian plate. Biostratigraphy of the recovered core sections was studied based on different fossil groups, magnetostratigraphic data, and results of electric and gamma logs. The distinguished biostratigraphic units of a high correlation potential add details to former stratigraphic schemes and characterize a series of isochronous levels traceable within and outside the West Siberian plate.

(2) Formerly, the Paleocene–Eocene sedimentation in the study region was regarded as uninterrupted, but we established several stratigraphic hiatuses, which may be related to displacements of basement blocks and eustatic fluctuations. The partial erosion of sediments took place in periods, when sedimentation regime in the West Siberian inner sea changed under influence of oceanic water masses located northward and in the south.

(3) The nannoplankton assemblage characteristic of the Gan'kino Formation upper strata is discovered for the first time in West Siberia. Planktonic foraminifers of genera *Acarinina, Hantkenina*, and *Turborotalia* are found at several stratigraphic levels in association with radiolarians, organic-walled phytoplankton, and palynomorphs of higher plants. Based on concurrent

assemblages of different fossil groups, we clarified and substantiated age of the green clay member with fish remains. Accumulation environments of this member crowning the upper Lyulinvor Subformation are elucidated. The Bartonian age of most typical marine facies of the lower Tavda Subformation is approved.

(4) Erosion surfaces established at the base and top of the green clay member of the upper Lyulinvor Subformation bring forward a problem of rehabilitation of the former Nyurol'ka Formation.

(5) Sea transgressions of eustatic origin and greatest amplitude took place in the Ypresian, middle Lutetian, late Bartonian, and late Priabonian. The Ypresian and Lutetian transgressions advanced in counter manner from the Tethys and Arctic Ocean. The Arctic influence on sedimentation dominated in the Ypresian and the Tethyan one in the Lutetian, as it follows from biogeographic analysis of different fossil groups. Transgressions and sea currents were interrelated. Throughout the Eocene, the southwestern part of the plate was within zone of the warm southern stream via the Turgai strait (Akhmet'ev *et al.*, 2001). That is why the most thermophilic and diverse assemblages of siliceous and organic-walled microplankton were found in association with planktonic foraminifers at several levels of the Lyulinvor and Tavda formations of the Barabinsk structural-facies zone. The basin shoaling in the early Bartonian and at the end of the Lyulinvor time antedated the commencement of sedimentation of the Tavda period.

(6) Stages of a considerable basin shoaling (terminal Thanetian, late Lutetian–early Bartonian, and Priabonian) are recorded at the levels of mass abundance of prasinophytes and acritarchs, where abundance and diversity of dinocysts is simultaneously reduced and the offshore distance coefficients are low. Shoaling of the mid-Tavda time was accompanied by freshening in the near-surface water layer. This event is evident from appearance of thin-walled *Hystrichokolpoma*, abundant spores *Hydropteris indutus* and *Azolla*, and infrequent *Pediastrum* specimens.

(7) We managed to determine age and to establish limitations of two transgressive–regressive sedimentary cycles of the Tavda Formation. The upper one is reduced and represented by the transgressive half-cycle only.

(8) In organic-walled phytoplankton of the Tavda Formation, the increase of abundance and diversity of cavate cysts is accompanied by growing content of endemic taxa. This is a consequence of climatic zoning and progressing basin isolation (in regressive phases especially) from oceans. Freshening of surface waters was also favorable for intense development of prasinophytes and acritarchs.

(9) In all transgressive members of the studied sections, there is microplankton redeposited from older strata. Cretaceous and even Jurassic radiolarians have been encountered in gray opoka-like clays of the upper Lyulinvor Subformation, Paleocene dinocysts in the green clay member of the same subformation, and Paleocene–early Eocene phytoplankton in transgressive clay member of the Tavda Formation.

(10) In southern areas of West Siberia, climatic fluctuations depended on changeable connections between the inner sea and adjacent oceans. Important in paleoclimatic aspect are palynological data on distribution of Fagaceae pollen in Ypresian–Lutetian deposits, primarily of taxa from *Castanopsis, Castanea*, and *Quercus* groups, because their geographic ranges are now under strict climatic control. According to these data, climate of the terminal Ypresian–Lutetian period in the study region was, like everywhere in middle latitudes of the former USSR except for the Far East, subtropical, of the monsoon type, with precipitation peaks in summer seasons. In the Bartonian and Priabonian, when the West Siberian sea lost connections with the Arctic Ocean, climatic conditions resembled those of the present-day Mediterranean region, where the precipitation maximum is characteristic of the winter period.

ACKNOWLEDGMENTS

We are grateful to I.A. Goncharova and S.V. Popov (Paleontological Institute RAS) for identifications of mollusks from the Tavda Formation. The work was supported by the Russian Foundation for Basic Research, project nos. 00-05-64917, 01-05-64320, 00-15-98490 ("Leading Scientific Schools"), 02-05-65170, and 02-05-64787.

Reviewers I.A. Basov and V.S. Popov

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Stratigraphy and Geological Correlation, Vol. 12, No. 5, 2004, pp. 514–522. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 12, No. 5, 2004, pp. 87–95.
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The Lower Paleocene of the Kronotskii Peninsula, Eastern Kamchatka

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Received June 5, 2003

Abstract—Upper Cretaceous and Lower Paleocene deposits of the Kronotskii Peninsula and their foraminiferal and radiolarian faunas are considered. As is shown, marine sedimentation during the Late Cretaceous–Early Paleocene was continuous in the Kronotskii and other peninsulas of the eastern and western Kamchatka, in the Koryak Upland, and Sakhalin Island.

Key words: Kamchatka, foraminifers, radiolarians, plankton, benthos, Upper Cretaceous, Paleocene.

INTRODUCTION

The Kronotskii Peninsula is central one in the structural–facies zone of eastern Kamchatka peninsulas (Fig. 1). Like in the Kamchatskii Mys and Shipunskii peninsulas, there are widespread Upper Cretaceous and Paleogene deposits mainly represented by volcanic–pyroclastic and tuffaceous–sedimentary rocks up to 5000 m thick in total. An intricate tectonics in the area, metamorphic alterations and a poor paleontological characterization of sequences hamper correlation of sections in different structural–facies zones. "Significant difficulties in interpreting stratigraphic succession of eastern Kamchatka are caused by several factors, primarily by almost identical lithology of Paleogene and Upper Cretaceous deposits, their flyschoid character, tremendous thickness, local transitions between sedimentary and volcanogenic rocks, and by intense deformations in Paleogene and Miocene strata" (*Geology of the USSR*, 1964, p. 194).

An important role in elaboration of Cenozoic stratigraphic scales for the Far East belongs to foraminifers, which are used, along with traditional mollusks, in biostratigraphic studies of the region since the 1950s. Beginning from the 1960s, radiolarians, calcareous nannofossils, and diatoms also received attention of the researchers. In the middle of the past century, lower Paleocene successions with very diverse and abundant assemblages of benthic and planktonic foraminifers were distinguished among Cenozoic deposits of the Kronotskii Peninsula. In this work, foraminifers were studied in samples of A.M. Sadreev, a geologist from the Kamchatka Geological Department, who collected them during geological survey of 1961. Samples characterize the lower part of the Kubovaya Formation exposed in southeastern coastal areas of the peninsula, where it is composed of alternating black pillow basalts and variegated andesites, basaltic tuffs, gray tuffstones, tuffaceous siltstones, and tuff conglomerates (Fig. 1).

SUBDIVISION OF CRETACEOUS AND PALEOGENE DEPOSITS IN THE KRONOTSKII PENINSULA

The sequences of volcanic, pyroclastic and tuffaceous–sedimentary rocks were first defined in the peninsula as the Kronotskii Formation (Grechishkin, 1935; Dvali, 1955; and others). It was established that the formation encloses a stratigraphic unconformity and its constituting parts differ in lithology (Sadreev and Dolmatov, 1965). Based on these features, the formation was subdivided into several units. Its lower part up to 700 m thick was discriminated from the rest as the Mys Kamenistyi Formation of the Late Cretaceous age. The upper thickest part of the formation attributed to the Paleogene retained the former name, being ranked however as a group. The Kronotskii Group was divided in two parts: the lower Kubovaya Formation with two subformations 1300 and 1600 m thick and the upper Kozlovskaya Formation up to 1400 m thick (Fig. 2). According to Sadreev and Dolmatov (1965), the Mys Kamenistyi Formation is composed of alternating black pillow basalts, tuffs of basaltic andesites, gray tuffstones, tuffaceous siltstones, and tuff conglomerates. Foraminifers have not been found in these rocks probably because of a limited number of samples collected for micropaleontological analysis. Tuffstones from the middle part of the formation yield spores and pollen. In opinion of V.P. Solomanovskaya, composition of palynological spectra is consistent with the Late Cretaceous age assumed for this portion of the formation (Sadreev and Dolmatov, 1965). The formation is characterized by radiolarians of Coniacian–Campanian age in the lower part and of Late Cretaceous–Early Paleo-

†Deceased.

gene (Danian?) age in the upper part (single locality) (Tsukanov, 1991).

Kronotskii Group: Subdivision and Paleontology

As is mentioned, the Kronotskii Group resting upon the Mys Kamenistyi Formation is composed of volcanic rocks approximately 4500 m thick in total, being subdivided into the Kubovaya and Kozlovskaya formations (Fig. 2).

According to Sadreev and Dolmatov (1965), the Kubovaya Formation overlies with a hiatus the Mys Kamenistyi Formation and consists of alternating dark-colored tuff conglomerates, psephitic tuffs, bleached tuffstones, tuffaceous siltstones, and banded siliceous rocks of the jasper type. The formation is 2900 m thick.

The Kozlovskaya Formation composed of pillow basalt flows alternating with psammitic and psephitic basic tuffs rests with an insignificant hiatus upon the Kubovaya Formation. Tuffs dominate in the upper part of the formation that is approximately 1400 m thick.

The Kronotskii Group was primarily considered as corresponding in age to the entire Paleogene. Subsequently, lower Paleocene planktonic and benthic foraminifers were found at the base of the Kubovaya Formation (Serova, 1966, 1969) and Oligocene mollusks in its upper part (Nesvit, 1967). Later on, different parts of the Kronotskii Group were dated based on nannoplankton (Shcherbinina, 1997). The Kozlovskaya Formation is determined to be the Late Eocene in age. The Oligocene planktonic and benthic foraminifers are discovered in the Rakitinskaya Formation (Stupin *et al.*, 1998), which overlies transgressively and with angular unconformity different strata of the Kronotskii Group.

Despite the thorough studies carried out recently in the Kronotskii Peninsula and data on additional orthostratigraphic fossil groups, there are controversial opinions about age of stratigraphic units defined here. In this connection, A. S. Arsanov, one of the most outstanding experts in geology of Kamchatka wrote: "...the relatively small Kronotskii area is first one in Kamchatka that was most intensely studied, but the results are ordinary. Available stratigraphic schemes are conflicting, some intervals are lacking distinct paleontological characteristics, and geological maps do not depict natural peculiarities of the tectonic structure... The objective cause of the low efficiency of regional geological studies in the Kronotskii area is its extremely intricate structure (Arsanov, 1978, p. 183). His words sound truly until present.

The available data on various fossil groups seem to be adequate for paleontological substantiation of Paleogene stratigraphic scheme of the Kronotskii Peninsula. The Late Cretaceous age of the Mys Kamenistyi Formation is proven based on radiolarian and palynological assemblages. Planktonic and benthic foraminifers from the lower 100–150 m of the Kubovaya Formation allow these strata to be correlated with the lower Paleocene *Globorotalia angulata* Zone. The higher portion of the Kubovaya Formation yields calcareous nannofossils of the lower Eocene *Discoaster lodoensis* (CP 11) and middle Eocene *Reticulofenestra umbilica* (CP 14) zones.

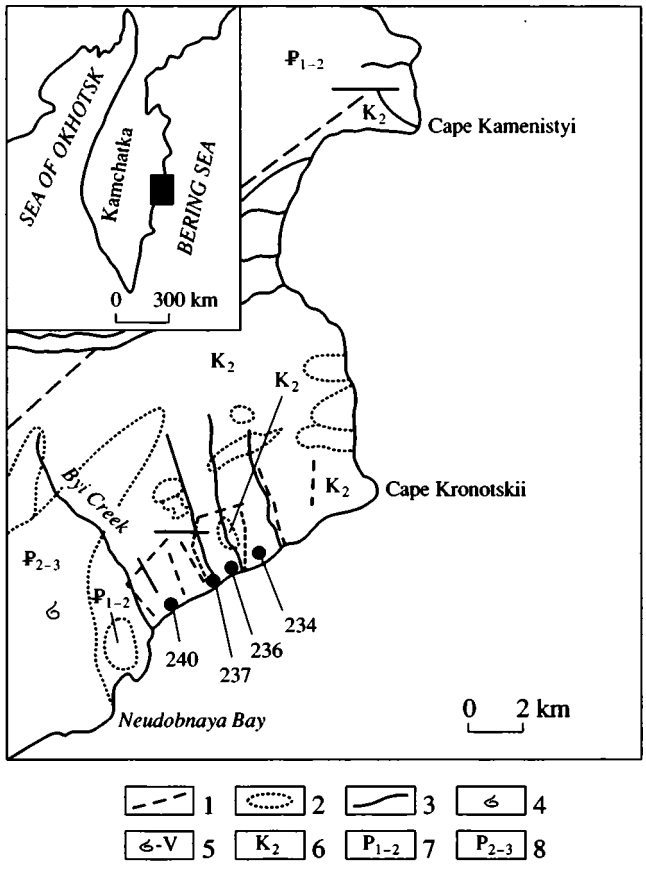

Fig. 1. Schematic geological structure of the Kronotskii Peninsula (compiled by A. M. Sadreev) and location of the study area (numbers in the figure correspond to sample numbers).

Nevertheless, Shapiro with colleagues (1996) cast doubts in their article, though without arguments, on presence of Maastrichtian–lower Paleocene and lower Eocene sediments, except for the Zone CP 11, in the peninsula. In addition to the Kronotskii Peninsula, the lower Eocene deposits with paleontological substantiation are known in other peninsulas of Eastern Kamchatka. Denying the Maastrichtian–early Paleocene age of the upper sequence of the Mys Kamenistyi Formation, the mentioned authors wrote: "two sequences are recognized in the Mys Kamenistyi Formation: the lower one with Coniacian–Campanian radiolarians and the upper sequence with Late Cretaceous–early Paleocene radiolarians determined in a single locality (Tsukanov, 1991). ...New finds of planktonic foraminifers and nannofossils from the upper sequence indicate its Eocene age (Beniamovskii *et al.*, 1992). According to additional determinations of planktonic foraminifers and nannoplankton, the lower, Kubovaya Formation of

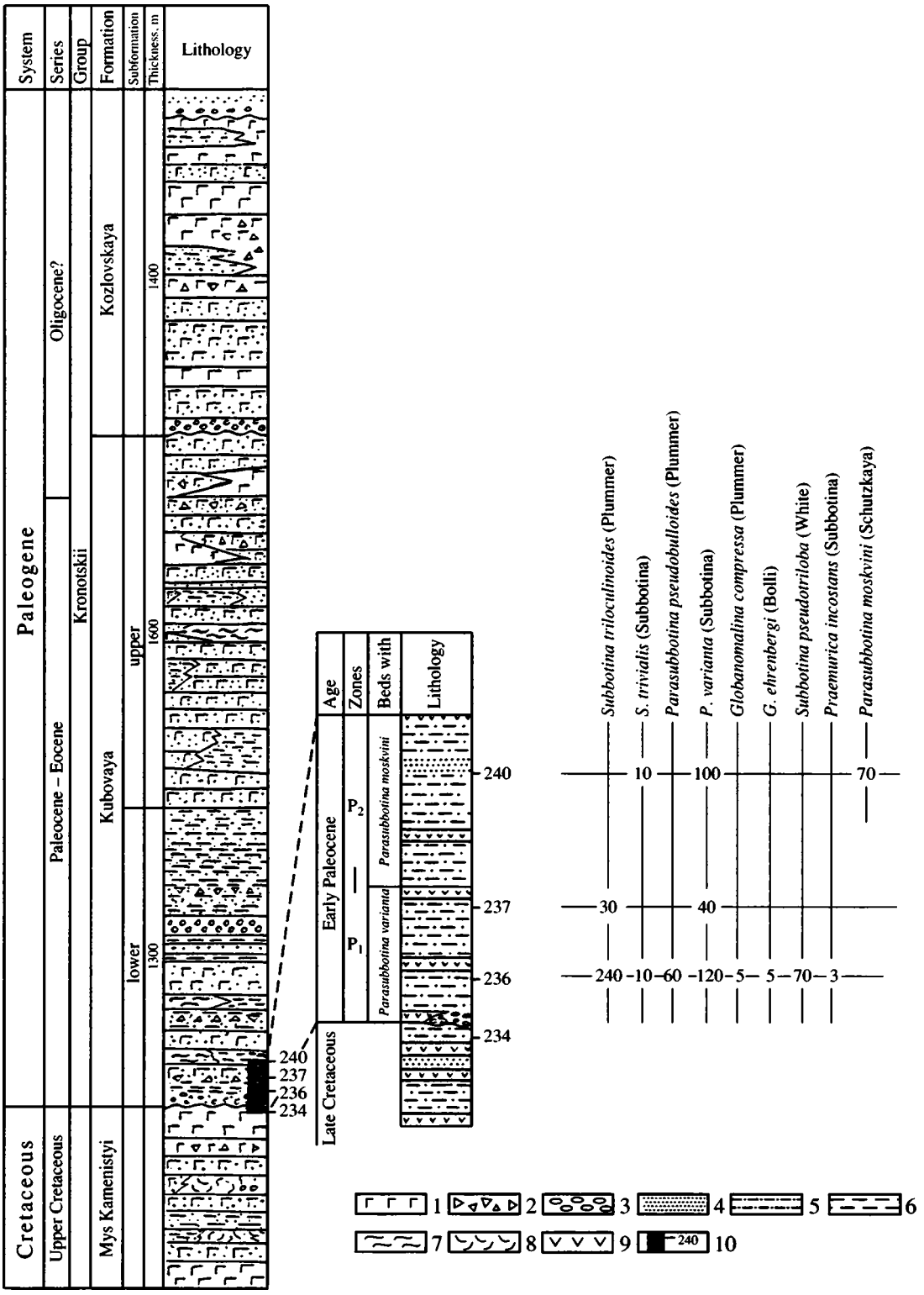

Fig. 2. Distribution of planktonic foraminifers in the lower part of the Kubovaya Formation (after Serova, 1966): (1) basalt; (2) tuff; (3) conglomerate; (4) sandstone; (5) siltstone; (6) argillite; (7) opoka; (8) melange; (9) tectonic contact; (10) sampled interval of the section.

the Kronotskii Group, which was previously attributed to the Paleocene (Serova, 1966), is referred partly to the uppermost lower and mainly to the middle Eocene (Shapiro *et al.*, 1996, p. 56).

Stratigraphic position of the Mys Kamenistyi and Kubovaya formations is a topic problem important for solving regional stratigraphic problems and for interpreting the geological development of Kamchatka and the entire Far East during the Cretaceous–Paleogene boundary period. As far as I know, none of the authors mentioned above is an expert in micropaleontology. Their article is lacking references, which could question presence of lower Paleocene strata bearing planktonic and benthic foraminiferal assemblages (Serova, 1966) in the Cretaceous–Paleogene boundary interval.

In this connection, it is reasonable to dwell once again on the micropaleontological characteristic of the lower Paleocene sediments in the Kronotskii Peninsula. Samples for micropaleontological analysis were collected by A. M. Sadreev from the lower 100–150 m of the lower Kubovaya Subformation. Sample 237, 236, and 240, were taken 50 m northeast of the Tret'e Ushchel'e mouth, at the Vtoroe Ushchel'e mouth, and 500 m north of the Bui Creek mouth, respectively. Sample 234 is from the Mys Kamenistyi Formation 1 km northeast of the Vtoroe Ushchel'e (Fig. 1).

The lower Paleocene deposits enclose a diverse assemblage (over 150 species) of planktonic and benthic foraminifers, described in my monograph and illustrated in plates (Serova, 1966). Stratigraphic ranges of most characteristic species of the Pacific and Atlantic provinces are presented in that book. For brevity sake, I analyze here only stratigraphic ranges of planktonic species from this assemblage.

The Mys Kamenistyi Formation

In Sample 234 from the Mys Kamenistyi Formation, foraminifers were not found. However, V.S. Vishnevskaya determined radiolarians *Orbiculiforma renillaeformis* Pess., *Amphibrachium* cf. *mucronatum* Lipman, *Prunobrachium(?) incisum* Kozlova, *Spongurum* sp., and *Prunopile* sp. in siliceous tuffaceous pelites from the upper sequence of this formation. In her opinion, these taxa indicate the late Maastrichtian–early Paleocene age of the formation (*Accretionary Tectonics...*, 1993, p. 242). The estimated age was confirmed in 2000 (Vishnevskaya, private communication).

The Kubovaya Formation

Samples from the basal part of the Kubovaya Formation (Fig. 1) yielded a diverse assemblage of benthic and planktonic foraminifers. This was first find of Paleogene planktonic foraminifers in the Far East, which was of principal importance for elaboration of the unified stratigraphic scheme accepted at the All-Union meeting of 1974 in Petropavlovsk-Kamchatskii, (*Resolutions...*, 1982). The assemblage of planktonic foraminifers consists of species widespread in lower Paleocene sediments of continents and oceans, although it is somewhat impoverished in taxonomic composition as compared with coeval assemblages from lower latitudes.

Two taxonomically different assemblages are distinguished in basal layers of the Kubovaya Formation (lower 100–150 m).[1] In the lower assemblage (samples 236 and 237), planktonic species are more abundant than benthic forms. These are *Subbotina triloculinoides* (Plummer), *S. trivialis* (Subbotina), *S. pseudotriloba* (White), *Parasubbotina pseudobulloides* (Plummer), *P. varianta* (Subbotina), *Globanomalina ehrenbergi* (Bolli), *G. compressa* (Plummer), and *Praemurica (Acarinina) inconstans* (Subbotina). Quantitatively dominant in the assemblage (up to 200–300 specimens in the standard weighed portion) are *Subbotina triloculinoides* (Plummer), *S. pseudotriloba* (White), and *Parasubbotina varianta* (Subbotina). Benthic foraminifers from the lower assemblage are less abundant being more diverse however (over 30 species of 14 genera). Dominant are anomalinids and cibicidids: *Cibicides praeventratumidus* (Massl.), *C. becki* Serova, *Anomalinoides nobilis* Brotz., and others. Sediments bearing the lower assemblage are defined as the *Parasubbotina varianta* Beds.

The upper foraminiferal assemblage is from fine-grained calcareous sandstones enriched in volcanogenic material and exposed in the interval of 100–150 m above the base of Kubovaya Formation (Fig. 1, Sample 240; Fig. 2). This assemblage also includes abundant planktonic foraminifers, which are represented by three species only. Two of them, i.e., *Parasubbotina varianta* (Subbotina) and *Subbotina trivialis* (Subbotina), are inherited from the lower assemblage. *Parasubbotina moskvini* (Schutzk.) is a new element in the assemblage. Benthic foraminifers are substantially more abundant and diverse than in the lower assemblage (over 70 species of 28 genera). In addition to rotaloid forms, buliminids, uvigerinids, and nodosariids are abundant and diverse in this case. Primitive rod-shaped agglutinated forms are accompanied by species belonging to *Haplophragmoides, Cyclammina, Ammobaculites*, and other more advanced genera. Sediments that yield the upper foraminiferal assemblage are defined as the *Parasubbotina moskvini* Beds (=Beds with *Globigerina varianta* and *G. moskvini* in Serova, 1966).

Planktonic foraminifers from the lower 100–150 m of the Kubovaya Formation are characteristic of the lower Paleocene *Globorotalia pseudobulloides, Acarinina uncinata*, and *Globorotalia angulata* zones (Fig. 2) recognizable worldwide (Serova, 1966; Kra-

[1] Previously (Serova, 1966; Krasheninnikov *et al.*, 1988), planktonic and benthic foraminifers were described in accordance with the classification accepted in *Fundamentals of Paleontology* (1959) and *Treatise on Invertebrate Paleontology* (Loeblich and Tappan, 1964). In this paper, generic names of planktonic foraminifers are given according to recent classification (Olsson *et al.*, 1992, 1999).

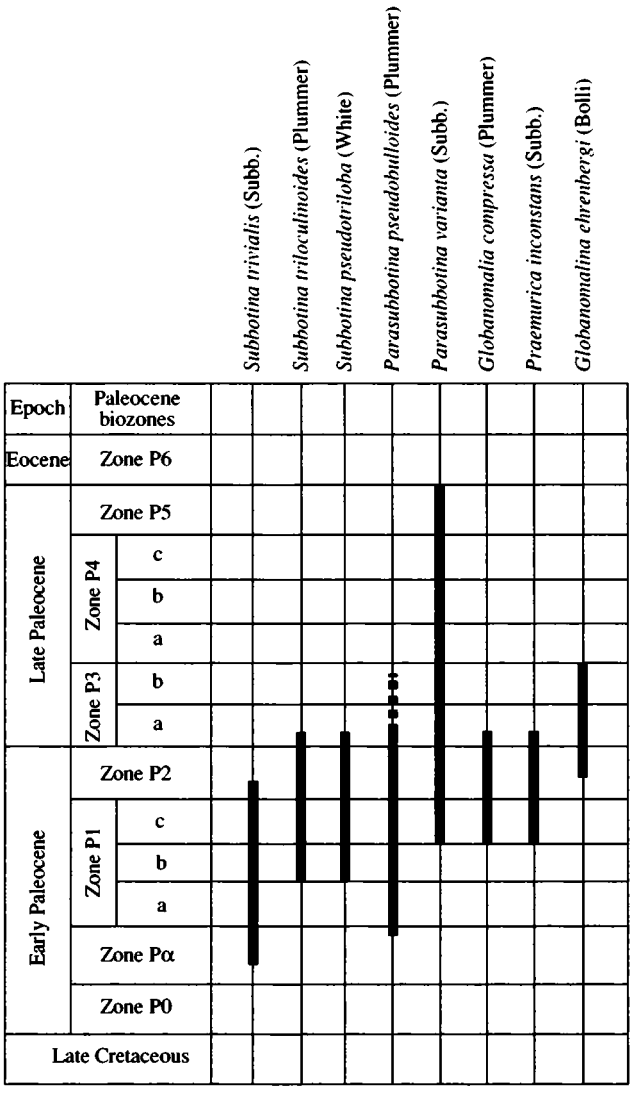

Fig. 3. Stratigraphic ranges of Paleocene planktonic foraminifer species (after Olsson *et al.*, 1999).

sheninnikov *et al.*, 1988; and others). Recent Paleocene zonal scale based on comprehensive analysis of available publications (Olsson *et al.*, 1999) is suitable for slightly more detailed subdivision of the interval into zones, boundaries between which correspond to first appearance and last occurrence levels of zonal species with due account for other biotic events. Authors of the scale determined ranges of Paleocene zonal and characteristic species, many of which are present in the Kubovaya assemblage of the Kronotskii Peninsula (Fig. 3).

Thus, the planktonic assemblage from the lower part of the Kubovaya Formation consists of species typical of Paleocene sediments of continents and oceans. *Subbotina trivialis, Parasubbotina pseudobulloides* and *Praemurica inconstans* present in the Kubovaya assemblage restrict the age of host sediments to the early Paleocene as it was established earlier (Serova, 1966; Krasheninnikov *et al.*, 1988).

Let us return to the paper by Shapiro *et al.* (1996), where presence of sediments older than the middle Eocene or initial early Eocene in the Kubovaya Formation section in the Kronotskii Peninsula is negated without any arguments. It should be emphasized that the early Paleocene age is established, based on planktonic foraminifers, only for its lower 100–150 m, not for the entire Kubovaya Formation (Serova, 1966). Inasmuch as the total thickness of the formation is up to 2900 m, it is logical to assume that remaining 2750 m could accumulate during the late Paleocene–Eocene time. In this case, age estimates will differ for different intervals of the formation section. This is perfectly demonstrated by data on distribution of nannofossils (Shcherbinina, 1997) in the Upper Cretaceous and Paleogene deposits of the eastern Kamchatka, the Kronotskii Peninsula included (Mys Kamenistyi Formation, Kubovaya formation with two subformations, and Kozlovskaya Formation).[2] Shcherbinina arrived at the conclusion that "...ages of indicated formations substantially differ from those determined previously, although the units themselves include different-age sequences" (Shcherbinina, 1997, p. 62) and that "...previously assumed stratigraphic relationships between the Mys Kamenistyi, Kubovaya, and Kozlovskaya formations are probably incorrect" (ibid, p. 64). For instance, nannoplankton species characteristic of the middle Eocene *Reticulofenestra umbilica* Zone (CP 14) were identified in samples collected from the exposure located between the Tret'e Ushchel'e and Bui creeks, where, according to Yu.B. Gladenkov and M.E. Boyarinova, the Mys Kamenistyi Formation is exposed. The same middle Eocene assemblage is determined from the lower Kubovaya Subformation exposed between the Vtoroe and Tret'e Ushchel'e creeks, from the upper Kubovaya Subformation south of the Baran'ya River and Izvilistyi Creek mouth, and from the lower part of the Kozlovskaya Formation exposed near the Vydrovaya River mouth.

At the same time, Shcherbinina (1997) determined the nannofossil assemblage of the Lower Eocene *Discoaster lodoensis* (CP 11) Zone in the lower Kubovaya Subformation near the eponymous river mouth and in its upper subformation south of the Ostryi Cape (Fig. 4). Neither lower Paleocene nannofossils (Shcherbinina, 1997) nor lower Paleocene foraminifers (Beniamovskii *et al.*, 1992) were found in section of the Kronotskii Peninsula because collection of Yu.B. Gladenkov and M.E. Boyarinova was most likely lacking samples from the basal part of the Kubovaya Formation, which yielded lower Paleocene benthic and planktonic foraminifers in the collection of A.M. Sadreev. Denying without arguments that lower Paleocene sediments with planktonic foraminifers are present at the

[2] Shcherbinina studied samples collected by M.E. Boyarinova and Yu.B. Gladenkov during field works in the Kronotskii Peninsula.

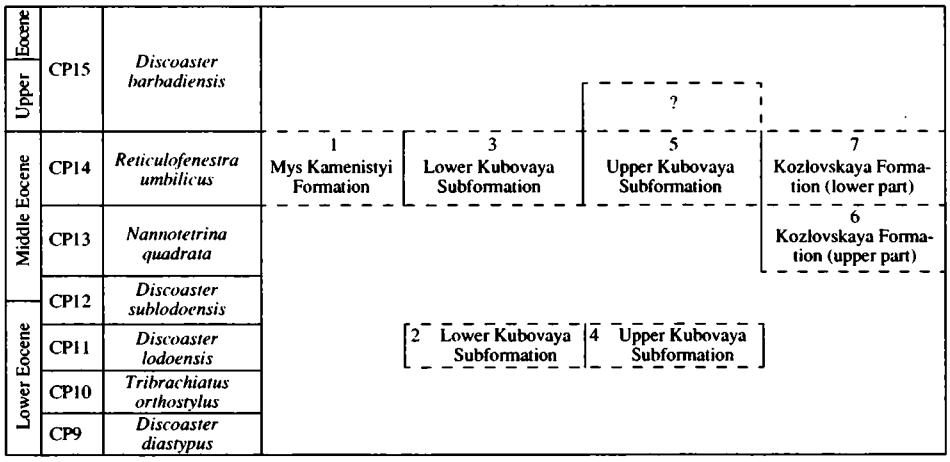

Fig. 4. Schematic correlation of sedimentary sections in the Kronotskii Peninsula based on nannofossils (after Shcherbinina, 1997): (1–7) exposures (1) between the Tret'e Ushchel'e and Bui creek mouth areas, (2) near Kubovaya River mouth, (3) between the Vtoroe and Tret'e Ushchel'e creeks, (4) south of the Cape Ostryi, (5) south of the Baran'ya River and Izvilistyi Creek mouth, (6) west of the Ushchel'e Creek, (7) near the Vydrovaya River mouth.

base of the Kubovaya Formation and Upper Cretaceous to Danian strata occur the upper part of the Mys Kamenistyi Formation, Shapiro *et al.* (1996) arrive themselves at the erroneous result implying a large regional break in sedimentation during the Late Cretaceous, Paleocene, and Eocene in the Kronotskii Peninsula.

Noteworthy is also another fact mentioned in the paper by Shapiro *et al.* (1996). Referring to data of Khubunaya (1987) they wrote: "...the relatively stable and specific geochemical characteristics of basalts beginning from the uppermost Cretaceous (lower Kamenistyi Mys Formation) up the uppermost middle Eocene (Kozlovskaya Formation) represent a peculiar feature of the Kronotskii Peninsula section...This can be considered as argument against a sharp reorganizations in tectonic development of the area." (Shapiro *et al.*, 1996, p. 56).

Ignoring the aforementioned paleontological and geochemical data, the authors insist on two principal points. They state that (1) paleontologically substantiated Paleocene and greater part of the lower Eocene are missing from the Kronotskii Peninsula, and that (2) this stratigraphic interval corresponds in western Kamchatka and the Sredinnyi Ridge to the sedimentation break related to tectonic activity, although angular unconformity between the Cretaceous and lower Paleogene layers has never been registered immediately in sections. Shapiro *et al.* (1996) believe that in eastern Kamchatka a hiatus at this level is admissible in the Kronotskii Peninsula only, where paleontological substantiation of Maastrichtian, Paleocene, and partly of Eocene strata is unreliable in their opinion. Simultaneously, they contend that there is no sharp unconformity at the Cretaceous–Paleogene boundary, and that changes in lithology and rock deformation degree are insignificant throughout the Maastrichtian, Paleocene, and greater Eocene intervals (Shapiro *et al.*, 1996).

Regarding paleontological and geochemical data as "very unreliable," the authors in fact find themselves in a tight corner because they cannot find an answer to question why sedimentation developed differently in a single structural–facies zone?

Shapiro *et al.* (1996) assert that, in contrast to continuous Maastrichtian–upper Eocene succession of the Kamchatskii Mys Peninsula, a large hiatus between the Campanian and uppermost lower Eocene is recorded in the Kronotskii Peninsula. Trying to explain this "phenomenon," they assume two possibilities: (1) objective differences in Late Cretaceous–Early Cenozoic development of these areas do exist; (2) "... thrust-related displacements and largely volcanic composition of the section hamper discovery of its missing parts in the Kronotskii Peninsula" (Shapiro *et al.*, 1996, p. 57).

I can agree with the last assumption. Taking into consideration an intricate structure of the area composed of many tectonic blocks, Shapiro and his colleagues should revisit sections, samples from which yielded fossils considered above. I mean spores and pollen of the Maastrichtian–Danian affinity (Sadreev and Dolmatov, 1965), Upper Cretaceous–Danian radiolarians from the Kamenistyi Mys Formation (Tsukanov, 1991), and lower Paleocene planktonic and benthic foraminifers from the lower Kubovaya Formation (Serova, 1966; Krasheninnikov *et al.*, 1988).

These finds undoubtedly indicate that geological development of the Kronotskii and Kamchatskii Mys peninsulas was similar, as it should be in a single structural–facies zone, and hiatus between the Upper Cretaceous–lower Eocene hiatus assumed by Shapiro *et al.* (1996) does not exist.

CRETACEOUS–LOWER PALEOGENE BOUNDARY SEDIMENTS IN NEIGHBORING AREAS OF THE NORTH PACIFIC

In addition to the Kronotskii Peninsula, paleontologically substantiated Upper Cretaceous–Lower Paleogene sediments are established in the eastern coast of Karaginskii Island. N.M. Petrina determined in the collection of S.A. Mel'nikova the foraminiferal assemblage of the Maastrichtian–Danian age (Chekhovich *et al.*, 1990).

In the well-exposed section along the Lozhnoil'pinskii Cape coast in the Il'pinskii Peninsula, Upper Cretaceous–Lower Paleogene sediments are represented by the flyschoid Old Il'pinskii Formation that is 1000 m thick and yields the Maastrichtian–lower Paleocene assemblage of benthic foraminifers (Serova, 1987).

The Upper Cretaceous–Lower Paleogene sediments are also present in the Govena Peninsula, where they yield rare specimens of early Paleocene planktonic foraminifers *Subbotina* cf. *triloculinoides* (Plummer) and radiolarians *Amphipyndax stockei* Campbell et Clark var. *C* Vishn. and *A. pluosios* Foreman, stratigraphic range of which spans the upper Campanian–lower Paleocene interval (Basov and Vishnevskaya, 1997).

The continuous Cretaceous–Paleogene boundary sections are also known in neighboring areas of the northwestern and northern Pacific. For instance, near Fresno and Curn in the Californian coast of North America, they are represented by the Panoche Group and by Moreno and Lodo formations, which form a continuous succession (Martin, 1964). The upper part of the Panoche Group (Uhalde Formation) contains a diverse assemblage of agglutinated and secretory benthic and subordinate planktonic (globotruncanids and rugoglobigerins) foraminifers, which indicate the Maastrichtian age of host deposits.

The larger lower part of the Moreno Formation referred to the Maastrichtian yields an impoverished assemblage of benthic secretory species. Agglutinated and planktonic foraminifers are missing from these sediments and only the upper member of the Dos Palos Formation yields the lower Paleocene planktonic species *Globoconusa daubjergensis* (Bronnimann) and rare benthic foraminifers (Martin, 1964; Serova, 1987).

Similar situation is also characteristic of Cretaceous–Paleogene boundary sediments in Hokkaido (upper part of the Nemuro Group) and in the Krasnoyarka River and Shakhtnyi Creek areas of southeastern Sakhalin (Kalishevich *et al.*, 1981; Serova, 1987). In Cretaceous sediments (the 1st and 4th subformations of the Krasnoyarka Formation), Turenko (1966, 1983) determined diverse benthic and planktonic foraminifers, globotruncanids and rugoglobigerins included, occurring in association with ammonites. Siltstones and carbonate concretions in the upper, 5th subformation of the Krasnoyarka Formation contain abundant foraminifers, rare coral remains, brachiopods, bivalves, spores, and pollen of the Paleogene affinity.

The foraminiferal assemblage dominated by agglutinated forms, like in underlying Cretaceous sediments, includes rare representatives of lower Paleocene planktonic species, e.g., *Subbotina triloculinoides* (Subb.) and others. Based on the integrity of enclosed fossils, the upper (5th) subformation of the Krasnoyarka Formation, which is defined as the Sinegor'e Horizon in the regional stratigraphic scheme, is estimated to be the early Paleocene in age (Kalishevich *et al.*, 1981; Serova, 1987).

There are also other sections with paleontological remains indicating continuous sedimentation in the Sredinnyi Ridge of Kamchatka (Serova *et al.*, 1970), eastern Kamchatka, and Koryak Highland during the Cretaceous–Paleogene boundary period. At any rate, the already mentioned data imply that sedimentation in the eastern peninsulas and islands of the Kamchatka region was continuous during the Maastrichtian–early Paleocene.

I cannot accept as well the idea that Paleocene sediments are absent in the western Kamchatka peninsulas facing the Sea of Okhotsk. Characterizing the coastal section of the Babushkin Cape (Mainach section), Shapiro *et al.* (1996) contend that shales of the Upper Cretaceous Omgon Group are overlain with angular unconformity by the middle–upper Eocene Snatol–Kovachina strata, and that hiatus spans here the Paleocene, lower Eocene, and lower middle Eocene intervals. Similar relationships are characteristic, in their opinion, of the Mt. Uvuch section as well. Indeed, in both these sections, the Paleogene sequence rests with a distinct angular unconformity upon shales of the Omgon Group and is composed of shallow-water sediments with numerous conglomerate interbeds, which could result from local breaks in sedimentation. Nevertheless, plant remains sampled between conglomerate beds at the base of the Cenozoic section in the 1950s represent, in opinion of E.N. Kara-Murza "...a flora transitional from the Late Cretaceous to the Tertiary one" (Krishtofovich, 1947, p. 32).

The inferable Paleogene age of sediments resting upon the Omgon Group is consistent with finds of benthic foraminifers in the lower conglomerate member (*The Eocene...*, 1991, Fig. 10, Bed 1) and in the lower sandstone layers from the middle part of the Snatol Formation (*The Eocene...*, 1991, Fig. 10, Bed 4?). Taxa identified at this level are *Rhabdammina indivisa* (Brady, *Silicobathysiphon nodosariaformis* (Subb.), *Silicosigmoilina* cf. *californica* Cushm. et Church, *Haplophragmoides glabratus* (Cushm. et Waters), *Asanospira* cf. *excavata* (Cushm. et Waters), *A.* cf. *akkeshiensis* (Yoshida), *Glomospira corona* Cushm. et Jarvis, and *Recurvoides* sp. (*The Eocene...*, 1991).

The lower 40 m of Paleogene sediments of the Mt. Uvuch section yielded not numerous foraminifers of the *Rzehakina epigona* Zone distinguished in lower Paleocene deposits of the Koryak Upland (Serova, 1987; Serova *et al.*, 1989).

A similar assemblage of benthic foraminifers characterizes the Sinegor'e Horizon of southern Sakhalin. It includes species known from the Lower Paleogene only, although they associate with many forms inherited from Cretaceous assemblages. Important components of this assemblage are rare planktonic species *Subbotina triloculinoides* (Plummer), stratigraphic range of which corresponds to P1–P2 and P3 zones the lower and upper Paleocene, respectively (Olsson *et al.*, 1999).

CONCLUSION

(1) The lower part of the Kubovaya Formation in the Paleogene section of the Kronotskii Peninsula (Kronotskii Group) includes lower Paleocene sediments with planktonic and benthic foraminifers and radiolarians. Planktonic foraminifers from the lower Kubovaya Formation are typical of the lower Paleocene sediments worldwide. The sediments characterized by the lower Paleocene assemblage represent an analogue of the P1a–c and P2 biozones (Olsson *et al.*, 1999).

(2) Taking into consideration a significant thickness of the Paleogene section, the intricate tectonic structure, and insufficient sampling frequency, it is reasonable to assume presence of upper Paleocene and lower Eocene layers with paleontological remains within the lower part of the Kubovaya formation by analogy with the Paleogene section of the Kamchatskii Mys Peninsula (Beniamovskii *et al.*, 1992).

(3) Presence of paleontologically substantiated Upper Cretaceous and Paleocene–lower Eocene sediments disproves inference by Shapiro *et al.* (1996) about the regional Late Cretaceous–early Eocene break in sedimentation within the structural–facies zone of eastern Kamchatka peninsulas, the Kronotskii Peninsula included. The available data on neighboring areas of the North Pacific region indicate absence of a significant hiatus at the Cretaceous–Paleogene boundary in the Bering Sea and Sea of Okhotsk paleobiogeographic provinces.

ACKNOWLEDGMENTS

I am grateful to A.S. Alekseev, the head of the Laboratory of Protistology, Paleontological Institute RAS, for the review of manuscript and valuable comments and to V.S. Vishnevskaya from the Institute of the Lithosphere of Marginal Seas RAS for donated data on radiolarians.

Reviewers A. S. Alekseev and I. A. Basov

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Stratigraphy and Geological Correlation, Vol. 12, No. 5, 2004, pp. 523–533. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 12, No. 5, 2004, pp. 96–106.
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On the Marine Genesis and Pontian Age of Deposits from the Lyubimovka Reference Section in the Crimea

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Received May 13, 2003

Abstract—A comprehensive study of the Lyubimovka section (western coast of the Crimea Peninsula, in the vicinity of Sevastopol) is carried out. The section is adopted as stratotype of six and as hypostratotype of three lower and basal middle Pliocene horizons in the stratigraphic scheme of continental deposits in the Ukraine. The late Miocene (early Pontian) age and the lagoonal–marine genesis of most deposits, which represent a typical molasse at the foot of the Crimean Mountains, is substantiated based on paleomagnetic, lithologic, biostratigraphic, and geological data. The results obtained imply that horizons of the lower and basal middle Pliocene continental deposits, which are distinguished in the Ukrainian stratigraphic scheme, are invalid.

Key words: Stratigraphy, Neogene, Miocene, Pontian, Lithology, biostratigraphy, paleomagnetism, Crimea.

INTRODUCTION

In southern Ukraine, particularly, in the Crimean steppe area, a sequence of variegated deposits is distinguished above the lower Pontian or upper Sarmatian. The sequence has been studied since the late nineteenth century and described under names of continental red–brown clays (Sokolov, 1883), variegated rottenstones (Fedorovich, 1928), and others. The Scythian clays occurring above the Khaprovian alluvium in the northern part of the Sea of Azov region and were assigned to that sequence as well.

Muratov (1954) attributed deposits older than Scythian clays to the Tavr Formation represented, in his opinion, by continental proluvium–alluvium with red–brown paleosols. He regarded the formation as a continental analog of upper Pontian and Kimmerian deposits (Fig. 1).

Following Fedorovich, Slavin (1977, p. 1140) found "several horizons of marine deposits amidst continental red beds that was inconsistent with the idea of exclusively continental origin of the Tavr Formation." Based on early Pontian mollusks (determinations by L.A. Nevesskaya and V.N. Semenenko) found in sandstone and limestone beds, he referred the Tavr Formation as a whole to the lower Pontian.

Kozhevnikov and Naidina (1983) initiated a comprehensive study of Tavr deposits in 1962 on the western coast of the Crimea from the Village of Uchkuevka to Lake Kizyl-Yar. In deposits assigned by Muratov to the Tavr Formation, they distinguished the Tavr Formation proper in the southern part of the studied profile and the Beregovoe Formation in the northern part, which is colored red less intensely than the Tavr Formation. The Tavr Formation was subdivided into six subformations (Fig. 1). In their opinion, each subsequent subformation is in onlap relation with the preceding one. Based on brackish-water ostracodes "found in all the intervals of rather extended marine section" (Kozhevnikov and Naidina, 1983, p. 118), a lagoonal genesis was suggested to the Tavr and Beregovoe deposits. Based on malacofauna found in limestone beds of the lowermost Uchkuevka Subformation, the cited researchers inferred the early Pontian age for this subdivision (fauna was determined by T.G. Kitovani and L.A. Nevesskaya). Four overlying subformations were referred to the late Pontian, and the uppermost Peschanoe Subformation to the Azov Horizon of the Kimmerian, while the Beregovoe Formation was thought to be spanning the rest of the Kimmerian and the Kuyal'nik Horizon. Besides fossils from the Uchkuevka Subformation, there are no reliable biostratigraphic data suggesting age of the Tavr and Beregovoe formations. The mentioned *Anancus arvernensis* found in deposits of the Beregovoe Formation cannot substantiate the Kimmerian–Kuyal'nik age of this unit, since this taxon is of a greater stratigraphic range.

Veklich (1982) who studied deposits in the lowland Crimea during the 1960s distinguished the deposits in question as a continental red-bed formation. He believed that this formation includes subaerial and subaqueous facies, the former represented by paleosols and the latter by alluvial–proluvial sand, sandstone, pebble,

†Deceased.

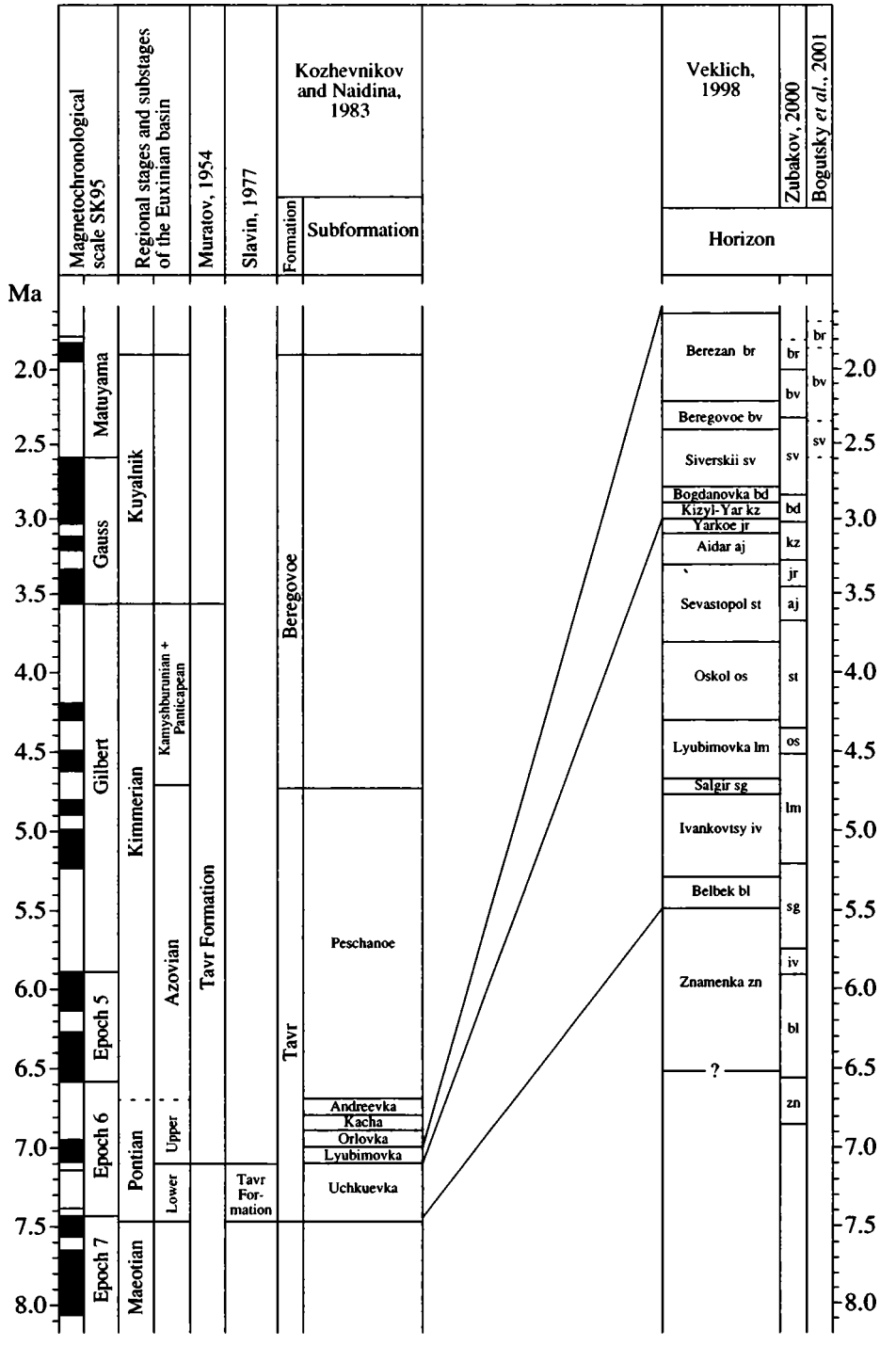

Fig. 1. Age of stratigraphic units of the Tavr Formation and horizons from "Stratigraphic Scheme of Pliocene Deposits in Ukraine" and their position in the magnetochronological scale.

conglomerate, and clay beds (Veklich and Sirenko, 1976; Veklich, 1982).

Peculiarities of paleosols and some palynological data were used to distinguished 16 stratigraphic horizons in the climatostratigraphic scheme of red-bed deposits in the Crimean western coast (Fig. 1), four of them were referred to the lower Pliocene (above the Novorossiisk Substage of the Pontian), six to the middle Pliocene, and the rest to the upper Pliocene (Veklich, 1982). In his work of 1998, Veklich estimated ages in million years for lower boundaries of the distinguished horizons (in Fig. 1, horizons of Veklich's scheme are arranged according to estimated dates).

Zubakov (2000), who also visited the Lyubimovka section, suggested the same succession though modified the boundary levels (Fig. 1).

Comparing the Lyubimovka section described by Kozhevnikov and Veklich, one may conclude that the Uchkuevka Subformation of the Tavr Formation shown in the scheme by Kozhevnikov corresponds to the interval from the Belbek Horizon to the top of the Yarkoe Horizon in the scheme by Veklich. The Lyubimovka Formation spans therewith the interval from the Kizyl-Yar Horizon to the Berezan Horizon included.

The stratigraphic scheme by Veklich is adopted by the Interdepartmental Stratigraphic Commission of the Ukraine in the rank of "Stratigraphic Scheme of Pliocene and Quaternary Deposits in Ukraine for Geological Maps of a New Generation" (1993; Bogutsky *et al.*, 2001).

When elaborating the Pliocene part of the scheme, Veklich regarded exposures of the Lyubimovka site as the main reference section that includes stratotypes of the Belbek, Salgir, Lyubimovka, Sevastopol, Yarkoe, and Kizyl-Yar horizons, hypostratotypes of the Ivankovtsy, Oskol, and Aidar horizons, and parastratotype of the Berezan Horizon.

This section is exposed in cliffs of the Crimea western coast south and north of the Belbek River mouth. Many researches studied it in different years. However, until present specialists are of different opinions with respect to genesis and age of deposits in the section, and we decided to study it comprehensively once more.

THE LYUBIMOVKA SECTION

Because of collapse events on shore, the section was exposed to a various extent in different years, and available descriptions of its structure differ therefore in details. In general, all the descriptions are fairly similar except for one work (Tretyak *et al.*, 1989), in which the section description has nothing to do with other descriptions, and paleomagnetic data published in that work useless unfortunately. The description by Veklich (Veklich and Sirenko, 1976; Veklich, 1982) is well recognizable. Moreover, when correlating deposits of the Lyubimovka section with stratigraphic horizons in the scheme by Veklich, we used a photograph of the cliff, he took from the sea, with his marks on it designating boundaries of the distinguished horizons.

The Lyubimovka section we studied north of the Belbek River mouth in 2001 is described below. The description includes results of field observations and microscopic study of rock structures and textures in thin sections. The section south of the Belbek River mouth, where the Belbek Horizon stratotype has been distinguished, is inaccessible for observations because of building activity in recent years. The cliffy coast north of the Belbek River valley is rising up along a distance of one kilometer northward to reach the maximum height about 37–38 m (Fig. 2a). All the deposits in the section gently dip northwestward (angle 3°–4°, azimuth 340°–345°), and youngest deposits are exposed therefore at the northern extremity. The cliff edges are unequally distant from the sea (Fig. 2b), and in places, where the cliffs are far away from the coastline (the latter is striking 5° N), lower parts of the section are covered with landslides.

The following beds (Fig. 3) are exposed in cliffs above the beach sands 2 m high (from the base upward; horizon names and indices from stratigraphic scheme by Veklich are given in brackets):

(1) Red to brick-red, massive, lumpy, silty clay (soil) with gliding planes; in the lower part it is mottled, brownish to yellow in color, with carbonate concretions. In thin sections, the clay matrix is of pelitomorphic to secondary colloform texture. In some areas, there are seen cellular and felty-fibrous textures (results of deuteric alteration of mica). Clay–ferruginous and pelitomorphic calcareous microconcretions are visible in matrix. Upper parts of the bed are saturated with humus substance (the Ivankovtsy Horizon, iv). Apparent thickness is 2.6 m.

(2) Light yellow, uniform, pelitomorphic clay, more silty in the upper part, microlaminated, slightly calcareous; carbonate is present as pelitomorphic calcite in the matrix and in authigenic crystalline and pelitomorphic structures or as inclusions of calcareous microorganisms (the Salgir Horizon, sg). The bed is 2.1 m thick.

(3) Alternating clay and light brown to brownish yellow silty clays, horizontally fine-laminated in the upper part; down the dip rocks are more reddish. The thin clays are pelitomorphic, with clear microlamination in separate areas, intensely humic, with abundant coaly detritus and signs of intense alterations (some areas are of the colloform texture, mica minerals are partially altered). Silty clays are lumpy, nonuniform, with pelitomorphic calcareous matrix and inclusions of carbonate pelitomorphic microconcretions. Remains of calcareous microorganisms are seen in thin sections throughout the bed (the Lyubimovka Horizon, lm). The bed is 2.5 m thick.

(4) Light, brownish gray platy limestone with sand and gravel lenses, yielding shells of early Pontian (list is given below) mollusks (the uppermost Lyubimovka Horizon, lm). The bed is 0.45–0.50 m thick.[1]

(5) Light yellowish gray calcareous silty clay that is darker, brownish gray in the lower part (intensely humic); a thin (10 cm) interlayer of brownish clay occurs in the middle of the bed (the lowermost Oskol Horizon, os). The bed is 8.7 m thick.

(6) Shingle with sand and gravel, cemented in places to conglomerate by carbonate matrix. As is seen in the northern part of the exposure, the bed bifurcates into three layers, being further completely replaced by

[1] Veklich and Sirenko (1976) considered this limestone as being an illuvial horizon of the overlying but completely washed out soil.

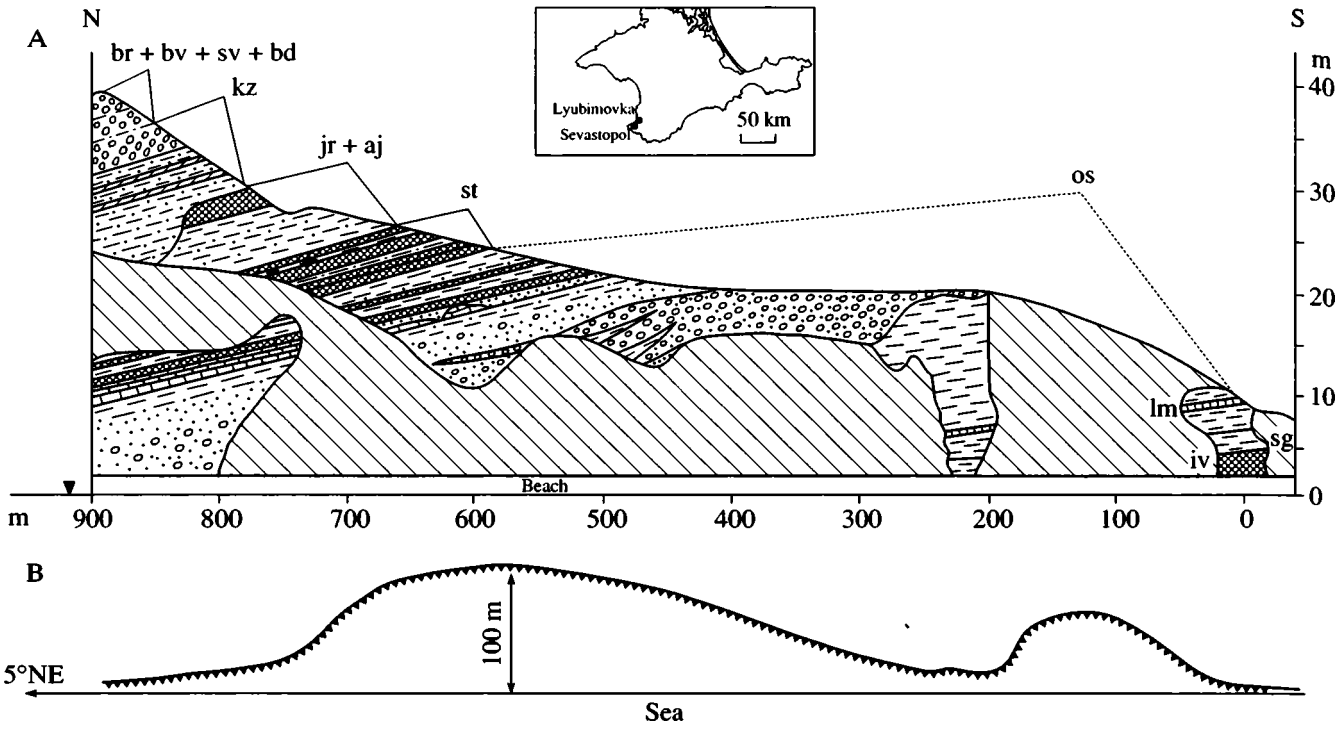

Fig. 2. Schematic geological section (A) and position of section edge relative to coastline (B), the Village of Lyubimovka (Crimean Pen.) (symbols as in Fig. 3).

sandy gravel and finally by sandy–clayey deposits (the Oskol Horizon, os). The bed is 6 m thick.

(7) Alternating gravel, light gray sand, brownish yellow silty clay, and greenish gray clay; proportion of clay material increases upward in the section. Silty clay is with pelitomorphic calcareous–clayey matrix and admixture of silty grains of quartz (up to 15%). Visible in places are areas of intense Fe-hydroxide and humus saturation with iron–manganese microconcretions. Clay is compact, laminated, with the colloform matrix texture, silt admixture, Fe–Mn microconcretions and hydroxides, carbonates, and plant detritus. A cream-colored pelitomorphic marl interlayer (up to 0.3 m) with a moderate admixture of silty quartz grains occurs 3.5–4 m above the bed base. Microorganic remains are detected in thin sections from interlayers of clay and silty clay (the Oskol Horizon, os). The bed is 6.5 m thick.

(8) Yellowish to reddish brown limestone comprising abundant gravel and pebbles, bearing shells of early Pontian mollusks; in the northern part of the exposure the bed is up 2 m thick. Amount of gravel and pebble considerably increases southward (opposite the bed dip), where limestone lenses appear and thickness decreases down to 0.8 m.

As is seen in thin sections, the organogenic compact silty limestone is grumous, ferruginate, composed of recrystallized calcite, shell detritus, and silty quartz grains; calcareous microorganisms are present in the rock as well. The calcite matrix is crystalloblastic and pelitomorphic. The bed encloses lenses and interlayers of clay, grayish yellow aleurite that is clayey, inhomogeneous, with pelitomorphic calcareous–clayey cement and unsorted grains of silt to fine-sand size. Characteristic of aleurite are abundant calcareous inclusions in the form of spots, pellets, concretions, and tubular root-shaped structures (hollow inside or filled with recrystallized calcite). Fragments of molluscan shells and calcareous microorganisms occur as well (the Oskol Horizon, os). The bed is 0.8–2 m thick.

(9) Silty clay (paleosol), reddish brown in the lower part, becoming yellowish brown in the upper part; matrix of the rock is fine-grained, pelitomorphic and colloform, with fragments of molluscan shells, well-rounded quartz grains, and with areas of intense impregnation by Fe-hydroxide (the Oskol Horizon, os). The bed is 0.2–0.3 m thick.

(10) Clay brown in the lower part, homogeneous, pelitomorphic to colloform, with impregnation areas of Fe-hydroxides; in the upper part it becomes pale yellow, pelitomorphic, more homogeneous, without silt admixture, calcareous, with ferruginate spots and inclusions of pelitomorphic calcareous concretions and rare foraminifers (the Oskol Horizon, os). The bed is 1.35–1.40 m thick.

(11) Reddish brown silty clay (paleosol), well distinguishable throughout section as a peculiar marker horizon; its matrix is colloform to cellular in texture.

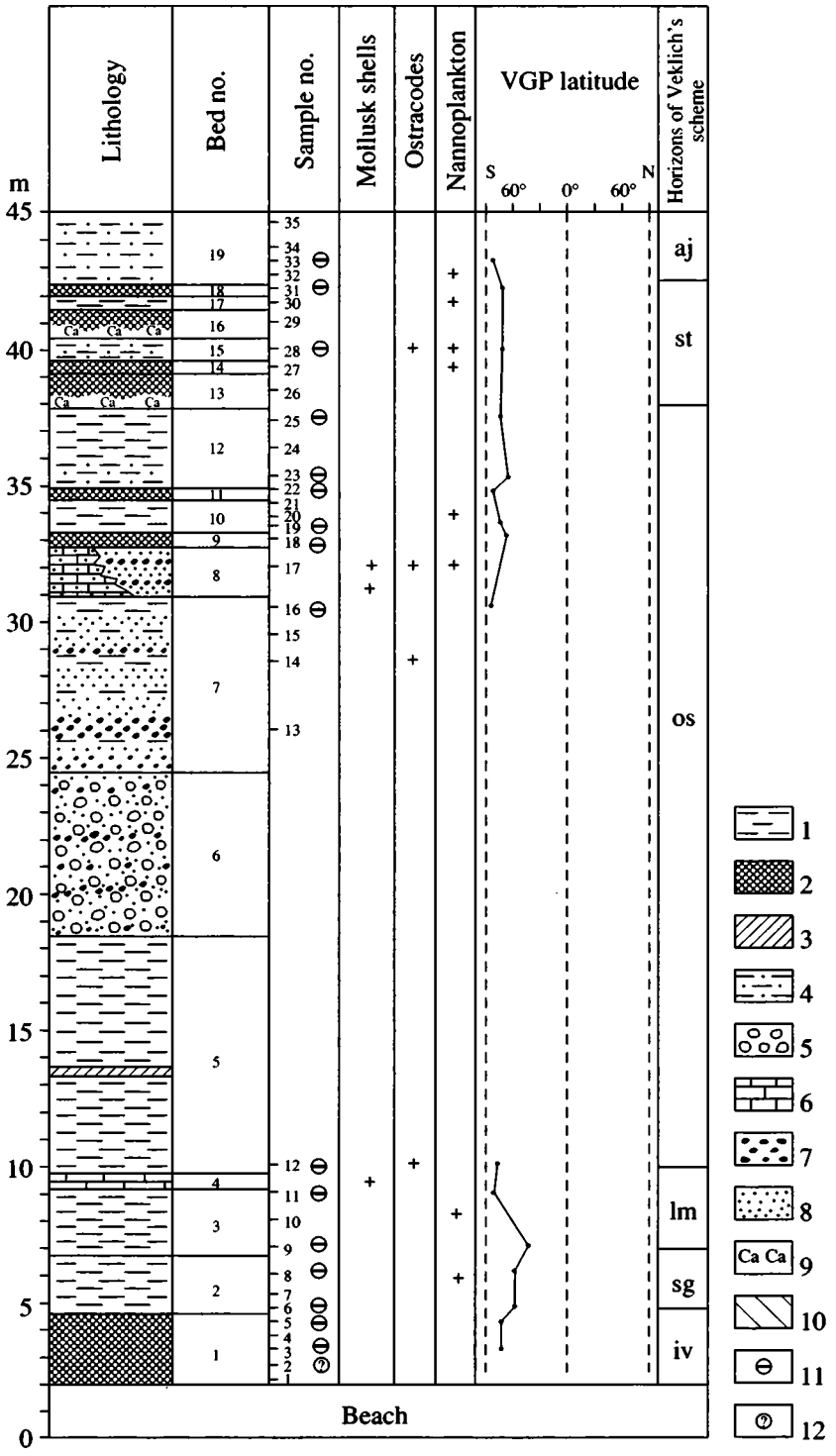

Fig. 3. Lithology, biostratigraphy and paleomagnetic characteristics of the Lyubimovka section: (1) light-colored, (2) red-brown, and (3) brown clays and silty clays; (4) silt and sandy clay; (5) shingle and conglomerate; (6) limestone; (7) gravel; (8) sand; (9) carbonate concretions; (10) talus and landslides; (11) reversed polarity; (12) polarity not established.

The bed comprises abundant sandy–silty admixture, intense impregnation with Fe-hydroxides, mica flakes with signs of deuteric alteration, and ferruginate altered glauconite represent characters of the bed (the Oskol Horizon, os). The bed is 0.25–0.30 m thick.

(12) Brownish yellow silt grading upward into brownish yellow clay; matrix in this calcareous–clayey silt is pelitomorphic, with Fe-hydroxide and calcareous microconcretions and fragments of molluscan shells. Clay is homogeneous, micro-laminated, pelitomorphic,

colloform in places, impregnated by Fe-hydroxides, with disseminated pelitomorphic and secondary recrystallized carbonate, bearing remains of calcareous microorganisms (the uppermost Oskol Horizon, os). The bed is 2.4–2.5 m thick.

(13) Bright red-brown silty clay (paleosol), mainly structureless, colloform, impregnated by Fe-hydroxides, bearing abundant coalified detritus and Fe-Mn microconcretions; rare tubular calcareous structures formed after roots were found in macerates. A light colored horizon (0.3 m) rich in carbonate concretions is distinguished near the bed base (the Sevastopol Horizon, st). The bed is 1.2 m thick.

(14) Silty red-brown, spotted, inhomogeneous, grumous clay with pelitomorphic matrix; the rock shows admixture of fine-grained sand, spots and concretions of pelitomorphic calcite (the Sevastopol Horizon, st). The bed is 0.4 m thick.

(15) Light gray, calcareous–clayey, inequigranular, inhomogeneous, poorly laminated silt with pelitomorphic calcareous–clayey cement, individual spots of Fe-hydroxides, dispersed grains of authigenic (?) silt-sized glauconite; coalified detritus and aggregates of calcareous algae were found in the bed. In the upper part of the bed, there is a 5-cm-thick interlayer of reddish silt (the Sevastopol Horizon, st). The bed is 0.9 m thick.

(16) Bright red-brown, grumous, inhomogeneous silty clay (soil) with pelitomorphic and structureless colloform matrix; carbonate abundant in matrix is present also in the form of pelitomorphic concretions. Intense Fe-hydroxide impregnation is observable. In the lower interval (0.3 m), the bed is rich in calcareous concretions (the Sevastopol Horizon, st). The bed is 1.25 m thick.

(17) Light gray calcareous silty clay (the Sevastopol Horizon, st); The bed is 0.35–0.40 m thick.

(18) Red brown silty clay (paleosol) similar to that of Bed 13 (roof of the Sevastopol Horizon, st). The bed is 0.40 m thick.

(19) Alternating light brown to brownish yellow clay and light yellow silt; clay is grumous, calcareous, pelitomorphic, with a minor admixture of silt and abundant pelitomorphic calcareous inclusions in the form of spots and pellets; remains of calcareous microorganisms are abundant in the bed. Silt is calcareous–clayey, inequigranular, similar to that of Bed 15. Remains of calcareous microorganisms are detected in thin sections throughout the bed (the Aidar Horizon, aj). The bed is 2 m thick.

It was impossible to study the section higher because of a large landslide. In Fig. 2, a part of section is illustrated based on descriptions by Veklich. It should be noted that the Aidar and Yarkoe (jr) horizons are about 7 to 8 m thick in total. The Yarkoe Horizon is represented in its uppermost part, from which Sample 36 is taken, by reddish brown, grumous, non-laminated silty clay with pelitomorphic calcareous–clayey matrix that encloses quartz grains of silt to fine-sand size, carbonate microconcretions, and calcareous microorganisms; there is an intense impregnation by Fe-hydroxides in some areas. It is well seen in the exposure that the Yarkoe Horizon top is truncated by argillaceous sand of the Kizyl-Yar (kz) Horizon (according to description by Veklich). The sand bed below overlying conglomerate is about 5 m thick near the southern end and more than 10 m thick (visible thickness) at the northern extremity.

The total thickness of the Kizyl-Yar, Bogdanovka (bd), Siverskii (sv), Beregovoe (bv), and Berezan (br) horizons distinguished by Veklich in the section above the non-truncated Yarkoe Horizon is up to 13–13.5 m.

CLAY MINERALS

Clay minerals are known to be the most sensitive indications of climatic conditions during sedimentation, and they have been carefully studied in this work. The mineral composition of clays from the described sequence is inherited in general from the rock complex that underwent erosion and redeposition. Undoubtedly, pedogenic processes did exert some influence, especially on minerals of the illite–smectite group, which are the most readily soluble.

The phase composition of clay minerals from the sequence is established using the X-ray diffraction method and diffractometer DRON-2 at the Laboratory of Physical Methods to Study Rock-Forming Minerals, Geological Institute RAS. A.L. Sokolova who analyzed diffractograms established a uniform phase composition of all the samples studied. Clay minerals are represented mainly by hydromica containing 5 to 20% of swelling (smectite) layers and by heterogeneous mixed-layered minerals such as illite, smectite, and, to a lesser extent, kaolinite. Sample 36 (the uppermost Yarkoe Horizon) contains an insignificant admixture of chlorite.

Distinctions between illite–smectite phases depend on the extent of biochemical reworking of originally smectite minerals in rocks during the pedogenic processes (influence of fulvic acids, soil bacteria, and others). Minerals of illite–smectite group are highly inhomogeneous in all the samples from fossil soils or sediments affected by pedogenic processes. These minerals produce indistinct (blurred) first basal reflections d (001) between 12 and 16 Å (often registered with difficulty) in diffractograms of air-dry preparations and between 18 and 20 Å if preparation is immersed in glycerol. They are detected in studied samples 1, 4, 22, 26, 27, 29, and 31 (Fig. 3). In other rocks not affected by pedogenic processes, diagnostics of the illite–smectite group are distinct small-angle basal reflections in the area of 13.8–14.6 Å in diffractograms of air-dry preparations and most distinct diffraction lines between 17.7–19 Å produced by glycerol-immersed preparations (samples 7, 10, 20, 24, 34, 36). The line near 19 Å was established only in two samples (28, 34). It charac-

terizes to a certain extent disintegration of illite–smectite indicative of rock transformation into soil. The degree of illite–smectite decomposition characterizes to some extent the intensity of pedogenesis. For instance, in samples 1 and 4 from the lower paleosol horizon of the studied section, illite–smectite appears to be completely disintegrated, because diffraction lines of the group are not registered in diffractograms. This is indirect evidence in favor of a highly humid climate, formation of plant organic matter, activation of fulvic acids and soil bacteria that resulted in dissolution of illite–smectite.

MOLLUSKS

Fedorovich (1928) was first to find shells of marine mollusks in the Lyubimovka section, but unfortunately, these fossils have not been identified reliably. Later on, Slavin, Kozhevnikov, and authors of this paper collected abundant paleontological remains identified by L.A. Nevesskaya, T.G. Kitovani, I.G. Taktakishvili, and V.N. Semenenko.

Shells of *Prosodacna* cf. *littoralis* (Eichw.), *P.* cf. *eichwaldi* (Andrus.), *Monodacna pseudocatillus* Barb., *Limnocardium* sp., and *Viviparus* sp. were found in Bed 4. Identified in Bed 8 are *Monodacna* ex gr. *novorossica* Sinz., *M. pseudocatillus* Barb., *M.* cf. *pseudocatillus* Barb., *Prosodacna* cf. *littoralis* Eichw., *P.* sp., *Congeria subcarinata* Desh., *C.* sp., *Dreisena* cf. *simplex* Andrus, *D.* sp., *Didacna* (?) sp., *Limnocardium* (?) sp., *Theodoxus* sp., *Melanopsis* sp., *Micromelania* sp. This assemblage is typical of the early Pontian and similar to "the fauna of the Odessan Limestone". It is the Odessan Limestone that Barbot-de-Marni (1869) took for the type unit the Pontian Stage.

OSTRACODES

N.N. Naidina identified an assemblage of marine ostracodes in collection of Kozhevnikov from beds 4 and 8, which includes "numerous species of genera *Tyrrhenocythere, Loxoconcha, Xestoleberis, Leptocythere, Caspiolla, Cyprideis*, and *Candona*. The most abundant taxon is *Tyrrhenocythere pontica* associated with rarer *Loxoconcha eichwaldi, Xestoleberis lutrae, X. plerique, X.* sp., *Leptocythere pontica, L. fragilis, L. dichathomica, Cyprideis torosa, Candona* sp. (juv.), *Caspiolla acronasuta*, and *Medicytherideis praepatoica*" (Kozhevnikov and Naidina, 1983, p. 112).

In our collection, V.A. Kovalenko found two forms of fresh-water ostracodes in Sample 12 (Bed 5) and one form in Sample 14 (Bed 7). Sample 17 from limestone of Bed 8 yielded a diverse assemblage of ostracodes (19 species) represented mainly by marine forms and some brackish-water (two) and fresh-water (four) species. In the clay lens enclosed in limestone (Sample 17a), fresh-water ostracodes (five of eight species) dominate over brackish-water forms. In Sample 28 (Bed 15), brackish-water and fresh-water forms occur (see the table).

Based on ecological affinity of ostracodes, it is inferred that sediments of beds 4, 8, and 15 were deposited in marine or lagoonal settings, which periodically experienced freshening. Freshening phases were local sometimes. For instance, when Bed 8 of coastal limestone accumulated, small creeks with fresh water entered the sea delivering the fine-grained terrigenous material and fresh-water ostracodes, because the ostracode assemblage in samples 17 and 17a, which were collected from the same level, close to each other though from of different lithologies (limestone and clay), are sharply different in composition. An important point is that brackish-water ostracodes were found in Bed 15 between fossil soils within the Sevastopol Horizon, as Veklich understood it. This is likely an indication of short-term lagoonal environments at the end of the Sevastopol time.

Because of a wide stratigraphic range, the found ostracode taxa are useless in terms of stratigraphic dating.

NANNOPLANKTON

According to S.A. Lyul'eva, samples 7, 10, 17, 20, 28, 32, and 35 yield the Late Cretaceous and Paleogene nannoplankton that is redeposited. The *in situ* forms were found in some samples however. These are *Dictyococcites antarcticus* Haq., smaller *Reticulofenestra* sp., *Calcidiscus macintyrei* (Braml. et Bukry), and *Braarudosphaera bigelowi* (Gran. et Braar.) from Sample 17; abundant *Perfocalcinella* ex gr. *fusiformis* Bona, rare *Calcidiscus macintyrei* (Braml. et Bukry), and *Dictyococcites antarcticus* Haq. from Sample 20; *Perfocalcinella* sp. and *Dictyococcites antarcticus* Haq. from Sample 28; rare *Perfocalcinella* sp. and *Calcidiscus macintyrei* (Braml. et Bukry) in Sample 32. Only rare specimens of *Perfocalcinella* sp. were found in samples 27, 30, and 36.

Nannoplankton was macerated from samples, if microorganisms were detected in their thin-sections. Cretaceous and Paleogene redeposited coccoliths were identified in macerates. Lyul'eva identified earlier *Dictyococcites antarcticus* Haq. in sediments of beds 2 and 3 and *Perfocalcinella* ex gr. *fusiformis* Bona in the upper part of Bed 7 of the same section (Semenenko *et al.*, 1999).

Nannoplankton taxa found are lacking zonal index species and cannot be used to date their host deposits. Nevertheless, they are indicative of marine sedimentaton environments during accumulation of beds 2, 3, 7 (the upper part), 8, 10, 14, 15, 17, 19, and of the Yarkoe Horizon (sample 36 from its upper part). *Perfocalcinella* forms abundant in Sample 20 suggest a phase of lowered salinity in the basin.

Ostracodes from deposits of the Lyubimovka section, marine (m), brackish-water (b) and fresh-water (f)

Taxons	Samples				
	12	14	17	17a	28
Euxinocythere praebosqueti Suz.			m		
E. multituberculata Liv.			m		
E. bosqueti Liv.			m		
Amnicythere propinqua Liv.			m		
A. aff. *olivina* Liv.			m		
Cyprideis torosa (Jones)			b	b	
Loxoconcha eichwaldi Liv., juv.			m		
L. laevatula (Liv.)			m		
Loxocorniculina djaffarovi (Schn.)			m		
L. *djaffarovi* (Schn.) var. *kertschinensis* Agal.			m		
Hemicytheria pontica Liv.			m		
H. ex gr. *aligera* St.			m		
H. praeazerbaidjanica Agal.			m		
Aurila. truncata Schn.			m		
Camptocypria acronasuta var.gracilis (Liv.)					b
C. acronasuta (Liv.)			b		
C. balcanica (Zal.)				b	
Cypria arma Schn.		f			
C. sinuata Schn.			f		
Caspiocypris ex gr. *candida* (Liv.)					b
C. filona (Liv.)					b
Candona kirgisica Mand.	f				
C. ex gr. *candida* (O. Muller)	f			f	
C. albicans (Br.)				f	f
C. subellipsoida (Scharapova)					f
C. suzini Sch.					f
Cypridopsis aff. *vassoevichi* Schn.			f		
Ilyocypris gibba (Rambdhor)				f	f
I. bradyi Sars.				f	f
Potomocypris longisetosa Bronstein			f		
P. aff. *plana* Schn.				f	
Chartocythere apatoica (Schweyer)				b	

GENESIS OF SEDIMENTS

Sorting and roundness of clastic material, lamination type, structure and texture of sediments, occurrence of calcium carbonate in the clay matrix (pelitomorphic, crystalloblastic), aggregated forms of Fe–Mn hydroxides, and extent of illite–smectite alteration were used as lithologic indicators elucidating formation environments of sediments. As a result, sediments are divided into three types deposited in basinal, shallow-water coastal (bar zone) and edaphic (paleosol) settings.

Characteristic of basinal sediments (beds 2, 3, 5, 12, 15, 17, and 19) are well sorted and rounded clastic material, fine horizontal and microscopic lamination often distinguishable in thin-sections only based on oriented unidirectional arrangement of mica flakes. Pelitomorphic calcite in clay matrix indicates, if present, sedimentation in saline seawater. Minerals of the illite–smectite group are slightly altered and retain their crystallographic diagnostic features. Single grains (plates, crusts) of authigenous glauconite were also found.

Shallow-water coastal sediments of bar zone (beds 4, 6, and 8) are composed of coarse sand, gravel and pebbles cemented by crystalloblastic and pelitomor-

phic calcite matrix. Pebbly and limestone beds (in places coquina) pinch out along the strike to be replaced by gravelly–sandy and clayey deposits. During subaerial exposition, they experienced influence of pedogenic processes as evident from tubular root-shaped calcite sheaths and incrustations near the top of Bed 8.

The genesis of sediments of the first and second types, which is inferred based on lithologic criteria, is consistent with findings of marine and brackish-water species of mollusks, ostracodes, and nannoplankton typical of marine lagoonal settings. Fresh-water ostracodes and *Perfocalcinella* forms abundant in some beds are indicative of recurrent freshening episodes in the basin.

Paleosols (beds 1, 9, 11, 13, 16, and 18) have colloform matrix. Sorting and roundness of their clastic fraction depend on the rock types in substrate. The intense impregnation by colloid–dispersed Fe–Mg hydroxides compounds is a result of crystallization of soil colloids. Carbonate segregations and concretions are formed in the course of calcite redistribution in soil horizons under influence of organic acids. Tubular root-shaped rhizocretions are characteristic. Illite–smectite is almost completely decomposed by soil bacteria and acids.

Beds 7, 10, 14 and sediments of the Yarkoe Horizon (Sample 36), which differ in their structure from sediments described above, are of special interest. Bed 7 and upper parts of the Yarkoe Horizon, which are composed of sandy–silty basinal sediments, accumulated at the time of shoaling and water freshening in the basin (confirming this are *Perfocalcinella* forms found at these levels). The sediments are affected by soil-forming processes as is evident from appearance of colloform matrix and intense Fe-hydroxide and humus impregnations at the top of beds. Bed 10 characterizes an opposite event: typical brown soil with the pelitomorphic–colloform matrix (Bed 9) was flooded during an inundation episode, when pale yellow to brownish yellow homogeneous fine clay was deposited on the soil surface. The distinct microlamination of that clay was established in thin-sections based on unidirectional orientation of mica scales. In inhomogeneous spotted red–brown clay of Bed 14 with abundant carbonate concretions, illite–smectite is almost completely decomposed. On the other hand, fine gradation of clastic material and pelitomorphic microlamellar structure of this clay suggest that sediment is a product of redeposition of underlying soil. To explain soil formation on limestones, it is reasonable to assume their deposition in a shallow-water basin that recurrently experienced exsiccation due to deposition of proluvial material derived from the adjacent Crimean Mountains. These sediments represent therefore a typical molasse (Semenenko, 2003).

PALEOMAGNETIC DATA

Remanent magnetization of samples collected by Veklich and Sirenko from the Lyubimovka section was studed by A.N. Tretyak since 1962 and by I.E. Volok since 1970s. Veklich (1982) published results of paleomagnetic studies for all the horizons of the section, however signs of rock magnetization in the lower Ivankovtsy and Sevastopol horizons are controversially indicated in his paper. Taking into account this ambiguity of paleomagnetic data reported by Veklich, we collected samples for paleomagnetic analysis from the lower and upper parts of the section to figure out paleomagnetic characteristics in these intervals (Fig. 3).

Beds 4 and 8 are deliberately referred to the lower Pontian based on molluscan fauna. Pontian deposits are known to show reversed polarity in general, and normal polarity is characteristic of their lowermost part only. Our intent was therefore to detect paleomagnetic reversal in the lower parts of the section and to establish how complete is the lower Pontian interval in the Lyubimovka section.

Samples for paleomagnetic analysis were collected from the exposure wall carefully cleaned. Three to eight cubes with facets 2 cm long were prepared from each sample. Paleomagnetic parameters were measured using magnetometer JR-4, the threshold sensibility 1×10^{-5} A/m. Since the sampled rocks are of different lithology, the measured values of their natural remanent magnetization (I_n^0) are different, $(0.85–78) \times 10^{-3}$ A/m, and the highest I_n^0 values are characteristic of deposits colored red. Two cubes from each sample were subjected to stepwise thermal demagnetization beginning from the room temperature to 700°C with intervals equal to 100°C. Characteristic of all the samples is unstable secondary magnetization completely disappearing at 200°C.

The component analysis showed that all the samples, except for Sample 2, recorded reversed polarity. We failed to determine the polarity sign for Sample 2 despite the twice-repeated full-scale test. The four cubes tested yielded different results, and Sample 2 was excluded from consideration. The mean latitudes of virtual geomagnetic pole (VGP) estimated based on demagnetization paths in the temperature interval 300–500°C are shown in Fig. 3.

The component analysis of paleomagnetic data showed that the lowermost Pontian deposits are missed from the studied section (rocks with normal polarity have not been found). Bed 1 with reversed polarity may be attributed to the lower Pontian or lower Maeotian. In the last case, we should suspect a considerable stratigraphic hiatus (not less than 0.5 m.y. long) between beds 1 and 2, but there is no geological data indicative of such a hiatus. Bed 1 is referred therefore to the early Pontian as well.

CONCLUSIONS

The analyzed lithology of deposits and found marine and brackish-water ostracodes, mollusks, and nannoplankton imply that sediments, exposed in the Lyubimovka section, accumulated in a shallow brackish-water basin that recurrently experienced freshening and occasional exsiccation, when soil horizons developed on marine sediments risen above sea level. Noteworthy are two major events related to a significant drop of the erosion base level. The first event is marked by deposition of a thick shingle layer (Bed 6), and second one resulted in substantial erosion of deposits of the Yarkoe Horizon.

Data considered above lead to the following age interpretation of deposits exposed in the Lyubimovka section. Shells of mollusks found in beds 4 and 8 evidence in favor of the early Pontian age of deposits in the section lower part. Judging from paleomagnetic data (exclusively reversed polarity of rocks), the lowermost Pontian deposits are missed from the section. Age of overlying deposits cannot be established based on biostratigraphic data. On the other hand, deposition of beds 12, 15, 17, 19, and the Yarkoe Horizon in a brackish-water basin (as well as brackish-water ostracodes found in Bed 15) allows us to attribute the respective interval of the section to the lower Pontian as well. This inference is consistent with conclusion of Kozhevnikov and Naidina (1983) who argued for the early Pontian age and lagoonal–marine genesis of the greater part of the Uchkuevka Subformation of the Tavr Formation. In terms of magnetochronological interpretation, the studied deposits correspond to Chron C3Br (7.432–7.09 Ma) in the scale SK95 (Pevzner *et al.*, 2003).

It is well known that an abrupt decrease in the Pontian basin area took place during the late Pontian in response to very extensive regression of that time. It was that regression that considerably lowered the erosion base level and resulted in substantial erosion of the Yarkoe Horizon and in superposition of overlying deposits. No reliable data are available to judge about age of deposits above the Yarkoe Horizon.

The thick shingle bed between two limestone horizons (within the Oskol Horizon after Veklich) may be indicative of the regression in the mid-early Pontian time, which is recorded in some sections of Ukraine and Moldavia (Bilinkis *et al.*, 1979).

The established early Pontian age of deposits within the interval from the Ivankovtsy through the Yarkoe horizons in nomenclature of Veklich testifies that horizons he distinguished in the Lyubimovka section belong to the Miocene and do not characterize the Pliocene continental deposits of Ukraine. The horizons are older than 7.1 Ma (the early–late Pontian boundary), whereas Veklich (1998) reckons them to be 3 to 5.3 Ma old. As deposits of lagoonal–marine genesis prevail in the section, the horizons of Veklich cannot be defined in lower Pliocene continental deposits of Ukraine. Unfortunately, they cannot be used for a detailed subdivision of the marine lower Pontian as well, since they are discriminated without clear criteria and are not recognizable even in neighboring sections. This is evident from different magnetic polarity signs assigned to presumably correlative horizons in different exposures (Veklich, 1982, Fig. 53). At the same time, deposits of regressive phase (thick shingle of Bed 6) may be of stratigraphic value, useful for subdivision of the lower Pontian in two parts in the local stratigraphic scheme.

ACKNOWLEDGMENTS

The work was supported by the Russian Foundation for Basic Research, project nos. 02-05-64126, 02-05-64428, and 03-05-64840.

Reviewers M.A. Akhmet'ev and S.M. Shik

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Stratigraphy and Geological Correlation, Vol. 12, No. 5, 2004, pp. 534–537. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 12, No. 5, 2004, pp. 107–110.
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On the Upper Induan and Lower Olenekian Substages of the Boreal Realm (Critical Comments to Article by Yu.D. Zakharov and Co-authors)

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Received April 7, 2003; in final form, July 9, 2003

Key words: stage, substage, boundary, zonal scale, section, ammonoids.

The journal **Stratigraphy and Geological Correlation** (2002, Vol. 10, No. 6) published the article by Yu.D. Zakharov, Ya. Shigeta, A. M. Popov, and others "Candidates for a Global Stratotype of the Induan–Olenekian Boundary (Lower Triassic) in Southern Primor'e." It is clear that selection of the GSSP between the Induan and Olenekian stages is an actual problem. Stratotypes of the stages are selected in different paleobiogeographic provinces, and this hampers solution of the problem. It is virtually impossible to find a section where boundary between those stages is substantiated based on paleontological data. Authors of the mentioned article attempted to solve the problem suggesting that sections in the southern Primor'e are proper candidates for the Induan–Olenekian boundary stratotype. In their attempt, they substantially revised the zonal scale of the upper Induan and lower Olenekian substages elaborated by Dagys and Ermakova (1993) for northeastern Asia.

Modification or perfection of biostratigraphic schemes usually involves new data or a thorough reexamination of available materials. New information could be obtained from previously unstudied sections, additional sampling of fossils, new methods, etc. What are grounds for such the cardinal changes in the zonal scale of the upper Induan–lower Olenekian substages of northeastern Asia in the case under discussion? Let us dwell on the zonal scheme of the upper Induan Substage (table). First I think that it is incorrect to speak of "frequently revised biozonation of the upper Induan" (Zakharov *et al.*, 2002, p. 584). It was nomenclature of subdivision, but not views on biozonation, which changed recurrently. Until 1983, in all biostratigraphic schemes, the Induan Stage corresponded to either a single or two zones: *Pachyoproptychites* (Popov, 1959), *Vavilovites* (Dagys *et al.*, 1979) *turgidus* and *Pachyoproptychites* (Vavilov, 1967), *strigatus* and *turgidus* (Archipov *et al.*, 1971), *turgidus* and *compressus* (Dagys, 1986). Moreover, the stage subdivision into the *turgidus* and *compressus* zones proposed by Vavilov was approved at the All-Union Meeting on Mesozoic Biostratigraphy of Siberia and the Far East in Novosibirsk, November of 1983. The experts in Triassic biostratigraphy doubted afterward a discrete character of assemblages from the *turgidus* and *compressus* zones and, thus, the approved subdivision. Unfortunately, many factual materials on the boundary interval have not been studied thoroughly by the meeting time. Based on preliminary data, it was assumed that index species of both zones occurs at the same stratigraphic level. To solve the problem, A.Yu. Egorov, A.G. Konstantinov, E.S. Sobolev, and myself carried out special field works in 1989. The study confirmed that section along the Berezovyi Creek, a left tributary of the Delin'ya River, is characterized by the complete succession of *Vavilovites* assemblages. It was shown also that section exposed along the nameless right tributary of the Delin'ya River corresponds to an upper part of the Berezovyi section, but not continues the latter as it was assumed by Vavilov (1982). Thus, fossil assemblages from the *turgidus* and *compressus* zones appeared to be coeval and doubts in their discrete character equitable. Subsequently, *Vavilovites* forms were thoroughly studied applying the ontogenetic and phylogenetic methods with due account for determination of species ranges and for directions and modes of the genus evolutionary transformation (Dagys and Ermakova, 1996; Ermakova, 2001). This provided a basis for the new zonation elaborated based on *Vavilovites* species (Dagys and Ermakova, 1993; Ermakova, 2002). In addition to the *Vavilovites* Beds, the new *Kingites? korostelevi* Zone was discriminated in the upper Induan Substage. Later on, the *Eovavilovites deliniensis* Zone was established at the base of the substage (Ermakova, 2002) largely owing to materials kindly donated by A.M. Trushchelev who collected them in 2000. Therefore, it is inappropriate to speak that authors of the Lower Triassic zonal scale (Dagys and Ermakova, 1993) did not have at their disposal "any significant and new factual data" (Zakharov *et al.*, 2002, p. 584). It is also incorrect to keep Dagys and Ermakova among authors of the modi-

Correlation of upper Indian–lower Olenekian Zonations

Stage	Substage	Northeastern Asia			Substage	Stage
		Dagys and Ermakova, 1993		"Dagys and Ermakova, 1993 (with modifications)" after Zakharov *et al.*, 2002		
Olenekian	lower	*Anawasatchites tardus*		*Anawasatchites tardus*	Ayakian	Olenekian (lower part)
		Lepiskites kolymensis		*Lepiskites kolymensis*		
		Hedenstroemia hedenstroemi		*Hedenstroemia hedenstroemi*		
Induan	upper	*Kingites? korostelevi*		?	Induan (upper part)	
		Vavilovites turgidus	*Vavilovites umbonatus*	*Tompoprotychites turgidus*		
			Vavilovites subtriangularis			
		Vavilovites sverdrupi		*Vavilovites sverdrupi*		

fied scheme as it is done (Zakharov *et al.*, 2002, p. 583). The modifications are usually shown in a separate column with indication of their authors.

What do Zakharov *et al.* (2002) suggest instead of the upper Induan zonation proposed by Dagys and Ermakova? The same that was proposed twenty years ago, i.e., the two-member subdivision of the *Vavilovites* Beds and the overlying empty interval with question mark. Moreover, *Vavilovites turgidus* is renamed without any arguments into *Tompoproptychites turgidus*. Maybe, it would be simpler to return to the scheme proposed by Popov (1959) and to set one zonal unit for the entire upper Induan Substage?

In the article under comments, the upper part of the Induan is shown in the correlation chart as the interval with question mark, which probably corresponds to the *Kingites? korostelevi* Zone, the upper unit of the Induan Stage in northeastern Asia. The arguments against validity of this zone are in essence as follows: "...all 16 investigated specimens of *Kingites korostelevi* Zakharov are obtained from the lower (Zakharov, 1978), but not upper Induan strata in interpretation by Dagys and Ermakova (1993)" (Zakharov *et al.*, 2002, p. 584). Indeed, Zakharov (1978) described 16 specimens, but he considered their host rocks at that time as the late Induan in age like did Dagys and Ermakova.

Let as consider the Burgagandzha Creek section, where these 16 specimens were found. The Permian sandstones are overlain there by "black shales with siliceous–calcareous concretions. The apparent thickness of the exposed sequence is 36 m. Talus below the cliff with exposed Triassic rocks yielded *Otoceras* sp. indet and *Vavilovites* sp. likely originating from different layers (boundary between layers with *Otoceras* and *Vavilovites* in the section is undeterminable)" (Zakharov, 1978, p. 33). Indeed, boundary between layers with *Otoceras* and *Vavilovites* is undeterminable here, because basal layers of the Lower Triassic are missed from this section that begins with the *Tompophiceras morphaeos* Zone. Because of displacement along the dextral strike-slip fault trending northeast, the Burgagandzha Creek section begins with the *Vavilovites turgidus* Zone, as it is confirmed by index species found in the exposure (Ermakova and Kutygin, 2000; Ermakova, 2002). The ammonoid assemblage collected higher in the section after a turf-covered interval includes *Kingites korostelevi, Episageceras dalailamae, Xenodiscus* cf. *subdemissium, Koninckites* cf. *kraffti, Ophiceras (Ophiceras)* sp. 1, *Ophiceras (Ophiceras)* sp. 2, *Glyptophiceras (Tompophiceras)* sp., and *Ophiceras (Lythophiceras)* sp. Ammonoids accompanying *Kingites korostelevi* are of a doubtful identification. For instance, specimen identified as *Koninckites* cf. *kraffti* Spath in Plate IV, figs. 7 and 8 (Zakharov, 1978) most likely belongs to the genus *Kingites*. *Koninckites* is the Tethyan genus whose representatives have never been found in the eastern part of the Boreal realm. In the area under consideration, the genus *Ophiceras* is represented by the only species *Ophiceras transitorium* occurring rarely. The Permian genus *Xenodiscus* is unknown from the Lower Triassic deposits. *Glyptophiceras* and *Lythophiceras* are also Tethyan taxa untypical of the Lower Triassic in northeastern Asia. Based on these doubtful determinations, Zakharov *et al.* (2002) revised age of deposits with *Kingites korostelevi* and attributed them to the lower Induan despite the *in situ* presence of underlying upper Induan and overlying lower Olenekian strata with *Hedenstroemia tscherski* (Popov). In principle, the section structure with the successive upper Induan, lower Induan, and lower Olenekian can theoretically be admitted in the case faults or capricious shear deformations, which requires unambiguous proofs. Any evidence of these deformations have not been found however in the course of joint field works with geologists of the Allakh-Yun' geological–prospecting expedition. Consequently, there are no grounds to consider the upper subdivision of the Induan Stage in the zonal scale of northeastern Asia as invalid. Besides the Burgagan-

dzha section, the *Kingites? korostelevi* Zone is traceable in some other sections of the eastern Verkhoyansk region, e.g., along the Onkuchakh Creek in the Yuge-Sakh River basin, Siren' Creek in the Allara-Sakh River basin, Berezovyi and right nameless creeks in the Delin'ya River basin, and along the Agylky River.

The upper Induan stratigraphic range of the genus *Kingites* is indirectly evident from its confinement to the *rotundatus–voluttus* Zone in the Salt Range (Waagen, 1985) and to the *Kingites discoidalis* Subzone in Arctic Canada (Tozer, 1994).

Similarly lame is criticism of the lower Olenekian zonation of northeastern Asia. Authors of the work under consideration doubted validity of the *Lepiskites kolymensis* Zone. Like in the previous case, it seems that they either are unfamiliar with or ignore the works published during the last two decades, those with revision of early Olenekian ammonoids of Siberia included (Dagys and Ermakova, 1990). Otherwise, they would pay attention to the fact that representatives of the genus *Hedenstroemia* in northeastern Asia indeed occur in association with *Lepiskites kolymensis*, but they belong to a new yet unnamed species. Unlike the earlier hedenstroemids, this form is larger, showing an intricate lobe line. This does not contradict the general trend in evolution of the genus *Hedenstroemia* toward a more complicated organization. Thus, the presence of *Hedenstroemia* sp. in the *Lepiskites kolymensis* Zone in no way violates the autonomous status of the latter. The age difference between *Hedenstroemia hedenstroemi* and *Lepiskites kolymensis* zones is confirmed by distribution of their index species in numerous sections from the Taimyr Peninsula to the Kolyma River. In the western part of the Boreal realm (Arctic Canada), the *Hedenstroemia hedenstroemi* Zone is defined at the base of the Olenekian Stage below the *Euflemingites romunderi* Zone (Tozer, 1994). Thus, the lower boundary of the Olenekian Stage in the Boreal realm is distinct and corresponds to the first appearance of *Hedenstroemia hedenstroemi.*

As is evident from the article "Candidates for a global stratotype of the Induan–Olenekian boundary (Lower Triassic) in southern Primor'e," its authors tend to unite two lower zones of the Olenekian Stage into a single unit. This would make somewhat easier the correlation between the lower Olenekian zonations elaborated for the Boreal realm and Primor'e region. As for the particular section proposed by authors for the Induan–Olenekian boundary stratotype (the Abrek Bay area), its correlation with Boreal sections is still ambiguous. In the Olenekian Stage stratotype, the stage lower boundary corresponds to the first appearance level of *Hedenstroemia hedenstroemi.* This level is easily recognizable in other regions of the Boreal realm as well. In many areas of the Tethyan realm, it is determined by first appearance of diverse hedenstroemids. Nevertheless, where is the "golden spike" placed in the boundary stratotype proposed by authors? It is at the base of bed with *Meekoceras boreale* Diener and *Ambites* sp. This bed yields however the ammonoid genera of a relatively wide stratigraphic range, which are studied to extent that is far from being sufficient. On the other hand, I am in perplexity with regard to the *Hedenstroemia bosphorensis* Zone range. The lower boundary of this unit is usually placed at the appearance level of its index species. In the considered section, the index species determined in open nomenclature appears in the middle of the zone. In this connection, it is unclear, in addition to question where the boundary should be drawn in the candidate sections for the boundary stratotype, whether there is any evidence indicating synchronism of the *Hedenstroemia hedenstroemi* Zone in the Boreal realm and the *Hedenstroemia bosphorensis* Zone in the Primor'e. Since authors want to unite two lower Olenekian zones of the boreal scale into a single unit, the *bosphorensis* Zone should correspond to the interval of *kolymensis* and *romunderi* zones. In such a case, the Primor'e region is lacking analogues of the lowermost *Hedenstroemia hedenstroemi* Zone. Another admissible interpretation is the condensed type of sections there, as they are widespread in the Tethyan realm (Tozer, 1971). Alternatively, ammonoids in Primor'e may be insufficiently preserved and inadequately studied. Because of this situation, it is unreasonable to accept the section, where position of boundary under consideration can be differently interpreted, for a candidate of the boundary stratotype. The lower boundary of the Olenekian Stage should be easily recognizable and unambiguously interpretable in the desirable candidate for the GSSP.

Reviewer A.S. Alekseev

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Stratigraphy and Geological Correlation, Vol. 12, No. 5, 2004, pp. 538–539. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 12, No. 5, 2004, pp. 111–112.
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To the Problem of the Induan–Olenekian Boundary (Response to the Critical Comments of S. P. Ermakova)

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Received April 24, 2003

Key words: Lower Triassic, Induan and Olenekian stages, Verkhoyansk region; Kolyma river upper course, southern Primor'e region, southeastern China.

The editorial board of the journal *Stratigraphy and Geological Correlation* familiarized me with the critical comments of S. P. Ermakova to our article (Zakharov *et al.*, 2002). I am grateful to S. P. Ermakova for her attention to the article and criticism. This reminded me a discussion with late A. S. Dagys in the journal *Albertiana* (Zakharov, 1994, 1995; Dagys, 1995) about the age of Lower Triassic Beds with *Kingoceras? korostelevi* in the Verkhoyansk region. The matter is that I am the author of the species *K.? korostelevi* Zakharov (Zakharov, 1978) and seemingly should be happy that precisely this taxon was chosen as an index species for the upper zone of the Olenekian Stage in Siberia (Dagys and Ermakova, 1993). Nevertheless, it seems that I was right when I opposed this choice, because I myself sampled fossils (although many years ago) and, thus, knew that, strictly speaking, beds with *K? korostelevi*, the species holotype included, do not correspond to the upper part of the Induan Stage in the Burgagandzha section. The top of the beds and base of the Olenekian Stage in this section, where Induan deposits are approximately of 400 m thick in total, are separated by a sandstone–siltstone sequence up to 200–300 m thick. I do not share optimism of my opponent with respect of the *Lepiskites kolymensis* Zone validity, because in the Ken'elichi section of the Kolyma River upper courses, I have found representatives of *L. kolymensis* (Popow) in association with numerous specimens of *Hedenstroemia mojsisovichi* Popow, the index species of the corresponding zone (Zakharov, 1978; Zakharov *et al.*, 2002).

The main purpose of our article (Zakharov *et al.*, 2002) is to search for sections suitable for candidates for the global stratotype of the Induan–Olenekian boundary, but not to criticism of the Lower Triassic biozonation used in the Boreal realm. I agree with every word of criticism said by S.P. Ermakova about the Abrek section in the southern Primor'e region, although new data on Induan and Olenekian conodont assemblages from this section appeared after publication of our article. In addition, the exact position of *Hedenstroemia* sp. indet. (Kiparisova, 1961) in the Abrek section is unclear, because the form is described from another collection. I do not agree however with opinion of my opponent relative the "golden spike." At present, it is important to have several candidates for the global stratotype of the Induan–Olenekian boundary. The number of proposed sections is however limited: (1) the Tri Kamnya Cape area in the Southern Primor'e region (Zakharov, 1994a, 1994b; 1996); (2) the Abrek Bay section in southern Primor'e (Zakharov, 2002; Zakharov *et al.*, 2000, 2002); (3) the West and North Pingdingshan sections in the Chaohu area, the Anhui Province of southeastern China (Tong *et al.* 2001a, 2001b, 2002a, 2002b). Carbonate and clay interbeds in the West and North Pingdingshan sections yield conodonts. They are particularly abundant in limestone interbeds (up to 200 specimens/kg), and the sections are most important despite a rare occurrence of well-preserved ammonoids there. As for ammonoids, the section yields late Changhsingian *Peuronodoceras attenuatum* Chao, Induan *Lithophiceras* sp, *Prionolobus hsuyuchieni* Chao, and *Ambites* sp. and early Olenekian *Arctoceras* aff. *lolouense* (Chao) Chao, *Euflemingites* cf. *tsotengensis* Chao, *Aspinites laevis* Chao, *Prosphingitoides* sp., *Juvenites orientalis* Chao, *Wasatchites* sp., and *Anasibirites* cf. *kwangsianus* Chao (determinations are mine).

As the chairperson of the International Working Group on the Induan–Olenekian Boundary (IOBWG), I am very interested in proposals of new candidates for the corresponding global stratotype, in addition to already available ones. This would make the work of the IOBWG more effective.

As is known, A. S. Dagys was an adherent of establishing the global stratotype of the Induan–Olenekian boundary in the Verkhoyansk region, although he has never proposed any candidate section. Basal layers of the Olenekian Stage (*Hedenstroemia mojsisovichi* Zone) with ammonoids and conodonts are easily trace-

able throughout the vast territory of northeastern Russia. On the other hand, a sequence immediately below them (a few tens of meters thick) is almost lacking macrofossils (conodonts have not been found) in the most complete sections examined (Burgagandzha, Ken'elichi, and others). In the Ken'elichi section, where Induan deposits are as thick in total as 500–600 m, the base of the *Hedenstroemia mojsisovichi* Zone is located 60 m above the layer with *Vavilovites kuluensis* Zakharov (Zakharov, 1978). At present, I am tending to consider *Vavilovites* and *Tompoproptychites*, which were previously defined in the rank of subgenera (Zakharov and Vavilov, 1976), as autonomous genera. At the same time, the criticism forces me to believe that an appropriate section will finally be found in the Verkhoyansk region as well. The International Working Group (IOBWG) expects also practical proposals from Jean Guex, the professor of the University of Lausanne, who studies materials from the Himalayas.

ACKNOWLEDGMENTS

I am grateful to the Presidium of the Far East Division, Russian Academy of Sciences for the support of the "Project of the Global Stratotype of the Induan–Olenekian Boundary (Based on Lower Triassic Sections in southeastern China and Primor'e)."

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